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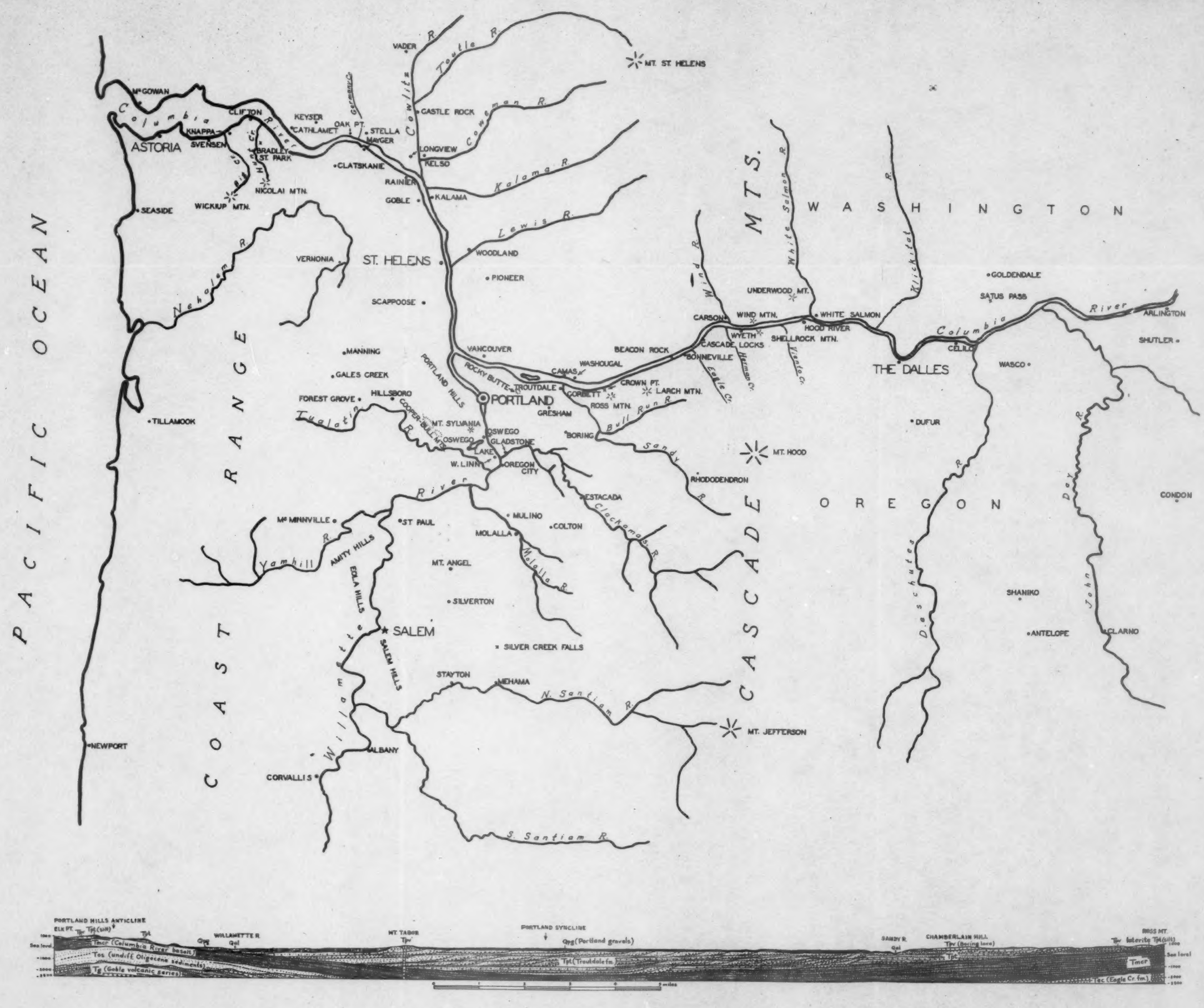

LOCALITY MAP OF THE LOWER COLUMBIA RIVER VALLEY, AND PORTLAND HILLS-ROSS MOUNTAIN (CROWN POINT) SECTION

(Section largely after Treasher)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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LATE CENOZOIC GEOLOGY OF THE LOWER COLUMBIA RIVER VALLEY, OREGON AND WASHINGTON

BY W. D. LOWRY AND E. M. BALDWIN

ABSTRACT

The Late Cenozoic formations in the lower Columbia River drainage basin constitute one of the best records of the terrestrial history of Western North America. The main events within the Cascade Mountains and westward to the Pacific Ocean, which culminated in the Cascadian Revolution, are:

1. Columbia River basaltic lavas were erupted from fissures during the middle Miocene upon a low-lying terrain of older Cenozoic sedimentary and volcanic rocks. Marine sediments of the Astoria formation were deposited just prior to extrusion in what is now the seaward part of the valley. The flows interfinger with beds which in general overlie the marine members of the Astoria formation.

2. Limonite beds and associated clastic sediments were deposited in depressions locally developed on the lava plain.

3. Additional lavas were extruded in the bog areas and perhaps on other parts of the lava plain.

4. While the Columbia River basaltic terrain lay near sea level, the uppermost basalt was extensively laterized to ferruginous bauxite. The Columbia River had established essentially its present course at or near the margin of the lava plain, but erosion prior to deformation was of little consequence.

5. Northerly trending open folds were produced in the basalt with those near the axis of the Cascade Range showing the greatest effect of the compression. Contemporaneously, anticlinal highs were modified by erosion and the lower part of the Troutdale formation was deposited in downwarps along the Columbia River while similar sediments were laid down in tributary valleys. Intercalated pyroclastics and flows herald the start of widespread Cascade volcanism. Basic lavas and pyroclastics were erupted from numerous vents including the major cones. All these events probably took place in the first half of the Pliocene.

6. Additional sediments were deposited as sea level rose in later Pliocene time. Most of the present valley areas were covered with fluviatile sediments and the rivers established their present courses on or from such fills. Most of the Cascade volcanic activity had apparently taken place before the Portland Hills silt member of the Troutdale was deposited in the late (?) Pliocene time.

7. Differential uplift began in late (?) Pliocene time and has continued to the present. The cutting of the Columbia River Gorge is the result of uplift which near the axis of the Cascade Range has been more than 2500 feet. Uplift in the Coast Range has been somewhat less.

8. Downcutting by the Columbia River during the (?) early Wisconsin, as a result of a eustatic lowering of sea level, amounted to approximately 600 feet. Deposition of the Portland gravels with their glacial erratics and contemporaneous sediments occurred with the ensuing rise in sea level.

9. Subsequent uplift in the Portland area has amounted to about 400 feet.

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Introduction

The formations present in the valley of the Columbia River constitute one of the most nearly complete records of the Late Cenozoic terrestrial history of western America. The relationships between the Late Cenozoic formations of the upper basin in central Oregon and Washington are fairly well understood, although a number of questions remain to be answered. In the lower basin of northwestern Oregon and southwestern Washington, the relationships between some of the Cenozoic formations, especially those of Miocene and younger age, need to be reconsidered.

Field studies during the past several years have disclosed much evidence bearing on these problems. Since late in 1943, the Oregon Department of Geology and Mineral Industries has discovered extensive deposits of ferruginous bauxite developed on the Miocene Columbia River basalt terrain in northwestern Oregon. These deposits are a good stratigraphic marker which has proved valuable in deciphering the geologic history.

The area discussed (Pl. 1) extends from the Cascade Mountains on the east to the Pacific Ocean. It includes the lower Columbia River Valley and the Willamette Valley as far south as Salem.

Acknowledgements

The writers wish to acknowledge the aid and encouragement given by Mr. F. W. Libbey, Director of the Oregon Department of Geology and Mineral Industries, under whose supervision much of the field work was undertaken. The manuscript was read critically by Mr. Libbey, Dr. R. L. Lupher, Mr. R. C. Treasher, Mr. Claire P. Holdredge, and Dr. E. L. Packard, all of whom gave information and offered many helpful suggestions. Information and suggestions were also given by Dr. I. S. Allison, Mr. Parke D. Snavely, Mr. Hollis M. Dole, Mr. S. E. Sargent, Mr. Paul W. Howell, and Miss Ellen James. A regional study that includes controversial topics naturally does not meet with universal agreement and the writers share responsibility for the ideas expressed. Much has been gained by discussion of the regional geology with many other geologists, and their aid toward a better understanding of the problem is appreciated.

Previous Work

Dr. Thomas Condon (1871) postulated an inland body of water within the Willamette Valley during the Pleistocene and called it the "Willamette Sound." Although the depth of water was not known, he believed that it reached nearly to Eugene as well as up the Columbia River to the Dalles and coincided in elevation with some of the marine terraces along the coast. I. A. Williams (1916) described deposits younger than the Columbia River basalt in his report on the Columbia River Gorge. J H. Bretz (1917) described the Satsop formation of southwestern Washington and discussed the relation of somewhat similar gravels which occur in the Columbia River

Gorge. He also suggested (1919) that a deep late Pleistocene submergence had affected the Columbia River Valley and wrote several papers (1925; 1928) describing the effects of Pleistocene floodwater, which he termed the Spokane flood. Any flood or floods in the upper Columbia River drainage would likewise have had their effect upon the lower part of its drainage.

E. T. Hodge (1938) presented a detailed discussion of the Columbia River Gorge and adjacent areas. He described the Troutdale, Rhododendron, and Cascan formations and also postulated a shift in the course of the Columbia River during the Pleistocene. He believed that lavas from Mt. Hood dammed the old channel and that, when the ponded water eventually spilled over, it cut the present channel farther north. I. S. Allison (1935) described Pleistocene gravel deposits and the origin and mode of emplacement of the glacial erratics in the Willamette Valley. He also proposed an alternate hypothesis of multiple floods and several stages of ponding behind ice jams to explain certain of the features of the Columbia River Valley. Treasher (1942) mapped the Portland area and discussed its geologic history.

Laterization of the Columbia River basalt and the thickness and distribution of a widespread overlying silt were discussed by Libbey, Lowry, and Mason (1945; 1946). Additional data on the distribution of the Troutdale gravel and overlying silt were presented by Wilkinson, Lowry, and Baldwin (1946) in their report on the St. Helens quadrangle. Other contributions will be discussed later.

STRATIGRAPHY

Miocene Formations

Columbia River basalt.—The Columbia River basalt is the most widespread Miocene formation in northwestern Oregon and the adjacent part of Washington. The term Columbia River basalt is used here to apply only to the Miocene flows as restricted by Merriam (1901) and Lindgren (1901). The basalt is interbedded with the marine Astoria formation of middle Miocene age in the lower part of the Columbia River Valley. Older formations, some of which are lower Miocene in age, are unconformable beneath the basalt and are not pertinent to this discussion. The attitude of the Columbia River basalt and the elevation and character of its surface is the key to post-basalt folding, faulting, and uplift which are embraced in the Cascadian Revolution.

LITHOLOGIC CHARACTER: The Columbia River basalt in northwestern Oregon consists of a series of dark gray to nearly black, fine-grained, basaltic lava flows with an average individual thickness of about 25–50 feet. Many of the flows are characterized by columnar jointing, whereas others are massive. Pillow structures, although not common, are present in the basalt in several places. The lower part of the basaltic section exposed along the Columbia River near Crown Point contains more than 50 feet of glassy breccia with some fragments several feet in diameter. In the Wickiup Mountain area about 12 miles southeast of Astoria near the mouth of the Columbia River, a thick series of angular basaltic flow breccias has been mapped as Columbia River basalt (Warren *et al.*, 1945).

The presence of a lighter gray, generally more inflated or vesicular variety or phase of the Columbia River basalt was pointed out to the writers by Donald Birch (personal communication). He has noted the occurrence of such lava in the Salem Hills and at several places in the Tualatin Valley.

The total thickness of the basalt flows is as much as 1000 feet in the Portland area. The Columbia River basalt thickens east of Portland and in the Columbia River Gorge where its maximum thickness is more than 2000 feet. In the Salem area, the maximum thickness is reported by Thayer (19[illegible]) to be about 400 feet. These figures probably represent nearly the entire original thickness as very little erosion has occurred in these areas where the laterite has been preserved at the top of the section. Because the lavas were extruded on an erosional surface and in most places have undergone erosion, the present thickness may differ greatly locally. In a few places steptoes of older rock may project above the surface of the basalt.

The Columbia River basalt formation in northwestern Oregon is not a continuous series of lava flows. In several places massive beds of conglomerate are present at or near the base of the section. Basaltic conglomerate interbeds occur in the basalt series on the south bank of

the Columbia River northwest of Mayger and across the river near Eufalla, Washington.

LIMONITIC INTERBEDS: At least one horizon of limonite deposits, mostly of bog origin, occurs in the upper part of the basalt series at a number of localities. The limonite, generally referred to as the Scappoose iron deposits, ranges in thickness from a few feet to nearly 19 feet. Associated with most deposits are minor amounts of fine clastic sediments. These limonite deposits do not form a continuous horizon, but deposits in addition to those near Scappoose are known near Oswego, in the Tualatin Valley, and in the Salem area. Although the limonite is not known to occur as far west as Bradley State Park on the south side of the Columbia River, the interbedded sandstone there and along Hunt Creek represents a break in volcanism which may be equivalent in age. Limonite deposits are not known along the Columbia River east of Scappoose, but weathered basaltic material in some of the basaltic sections there may have been developed at the same time.

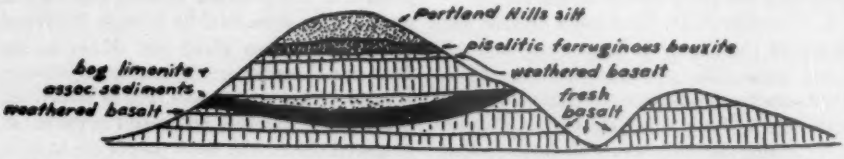


FIGURE 1.—IDEALIZED CROSS SECTION THROUGH THE COLUMBIA RIVER BASALT IN COLUMBIA COUNTY, OREGON

LATERITIC FERRUGINOUS BAUXITE: The lateritic ferruginous bauxite deposits which occur at the top of the uneroded basaltic section are much more widespread than the limonite deposits. They occur in an elongate area extending from a point south of Salem, Oregon, to a point just north of Longview, Washington. There is no genetic relationship between the limonitic and the bauxitic deposits (Fig. 1). At one locality a short distance northwest of Scappoose in Columbia County, 75 feet of weathered basalt, which Hotz (1942) believes includes three separate flows, separates the limonite from the laterite as suggested in the figure.

The ferruginous bauxite deposits in northwestern Oregon are at the top of the lateritic section which is in places more than 50 feet thick. The bauxitic horizon, with a silica content of 10 per cent or less, ranges in thickness from a few feet to as much as 20 feet, largely as the result of erosion subsequent to the laterization and prior to the deposition of the massive silt which overlies many of the deposits. Most of the bauxite is reddish-brown and has a nearly uniform chemical composition. The upper part is characterized by a pisolitic texture; the lower horizon may differ in texture from nodular to earthy. The widespread effect of this intensive laterization indicates that the Columbia River basalt underwent a prolonged period of weathering during which erosion was negligible. Presumably the basalt terrain lay near sea level while laterization was taking place. If so, this horizon is very valuable in determining subsequent movement relative to sea level.

STRUCTURE: The attitude of the lateritic deposits, as well as the limonitic interbeds, corresponds to that of the basalt which in most places dips less than 10°. Most of the deposits dip only a few degrees. Reliable dips in the basalt are difficult to obtain because of the thick soil mantle; in most places where the attitude is best shown, only the general direction of dip can be taken.

The topographic distribution of the basalt ranges from as much as 1200 feet below sea level in the Tualatin Valley and Portland area to more than 3000 feet above sea level on Nicolai Mountain in Clatsop County. The Columbia River basalt in northwestern Oregon has been folded into broad synclines and anticlines. The Tualatin Valley, the Portland syncline, the synclinal valley between Mt. Angel and the Eola and Amity Hills, and the Stayton basin are partly filled synclines in the basalt; and the Portland Hills and the Cooper Mountain-Bull Mountain ridge are anticlines. The basalt has been faulted in a number of places but the displacement along most is not great.

Most faults, as well as the axes of folds, trend northwesterly.

The Columbia River basaltic lavas were erupted from relatively narrow fissures whose former presence is indicated by columnar jointed dikes.

The basaltic terrain is characterized by a youthful topography. Many of the divides are quite broad and relatively undissected, and some still retain a partial cover of the laterite developed on the original basaltic surface. The youthful character is also shown by the steep gradients of the streams draining the terrain.

AGE: Weaver (1937) has pointed out that the Columbia River basalt on the Washington side of the Columbia River contains interbedded marine sandstone with fossils of the Astoria formation of middle Miocene age. The contemporaneous age of the Columbia River basalt and Astoria formation is also indicated by the interbedded nature of Astoria sandstone and basaltic lavas on the south side of the river between Bradley State Park and Astoria. Elsewhere in northwestern Oregon, the Miocene age of the basalt is indicated by its unconformable relationship with the underlying marine Scappoose formation of upper Oligocene or lower Miocene age, and its unconformable relationship wherever seen in contact with the overlying Troutdale formation of Pliocene age. In the Columbia River Gorge, the Columbia River basalt lies on an erosional surface cut in sediments of the Eagle Creek formation. The fossil flora of the Eagle Creek formation has been assigned by Chaney (1944) to the lower Miocene. From the gorge the Columbia River basalt can be traced eastward along the river and it tributaries into central Oregon, where, as in Picture Gorge, it lies on an erosion surface with a relief of several hundred feet. The erosion surface was developed on the John Day formation of late Oligocene or early Miocene age. The upper age limit of the basalt is indicated by the late Miocene age of the overlying conformable Mascall formation.

Columbia River basalts in the vicinity of Salem and Stayton, Oregon, were named the Stayton lavas by Thayer (1939), as the Columbia River basalt had not yet been traced southward into the area. However, it now seems advisable to drop the name Stayton lavas in preference to Columbia River basalt and, except for the intervening alluvial fill of the Willamette River, exposures would be continuous. Andesitic or less basic basaltic lavas, which occur south and east of Stayton, appear to belong to the series (Thayer, 1939).

Astoria formation.—The Astoria formation, near the mouth of the Columbia River in the vicinity of Astoria, Oregon, consists of a relatively thick series of Miocene marine sediments which Howe (1926) divided into a lower sandstone, the well-known shales, and an upper sandstone member. The shales have been intruded by and are below the horizon of the interbedded Columbia River basalt. Several miles east of Astoria in the Svensen and Cathlamet quadrangles, none of the characteristic dark marine shale of the Astoria formation was found and the Astoria sediments both above and below the basalt are yellow-stained friable sandstone, the upper member of which may not be entirely marine.

Only that part of the Astoria formation interbedded with and resting upon the Columbia River basalt, and designated the upper member of the Astoria formation, is pertinent here.

UPPER ASTORIA SANDSTONE: At least 500 feet of sediments overlie the Columbia River basalt in the vicinity of Clifton and Brownsmead, Oregon. Washburne (1914) first noted the probable upper Miocene age of the beds near Clifton. The basal member of this series is a yellow, massive, friable sandstone that is exposed along the highway west of Bradley State Park, along Gnat Creek, and along the road east of Knappa. This series is overlain by platy carbonaceous sandstone and sandy shale which crops out along the railroad from Clifton westward nearly to Brownsmead. Along the Clifton side, the absence of the basal sands above the basalt may indicate a slight disconformity. The section is generally conformable in dip with the underlying basalt, and the basal sands are very similar to sands which are interbedded with the basalt along Hunt Creek and also those exposed on the high way loops just east of Bradley State Park and to the sand lenses within the Astoria formation farther west. The interbedding of part of this series with the basalt and the unconformable relationship of these beds with overlying horizontal sediments which are tentatively correlated with the Troutdale formation, suggest that the sediments of this series are related to

and perhaps belong to the uppermost part of the Astoria formation.

The writers traced the series of basalt flows westward from Bradley State Park to Big Creek where they are no longer well-defined flows but that trends northwesterly from Mayger across the Columbia River to a point west of Stella, Washington. These tuffaceous conglomerates and sands have been thoroughly altered to somewhat bauxitic clays. They are exposed at

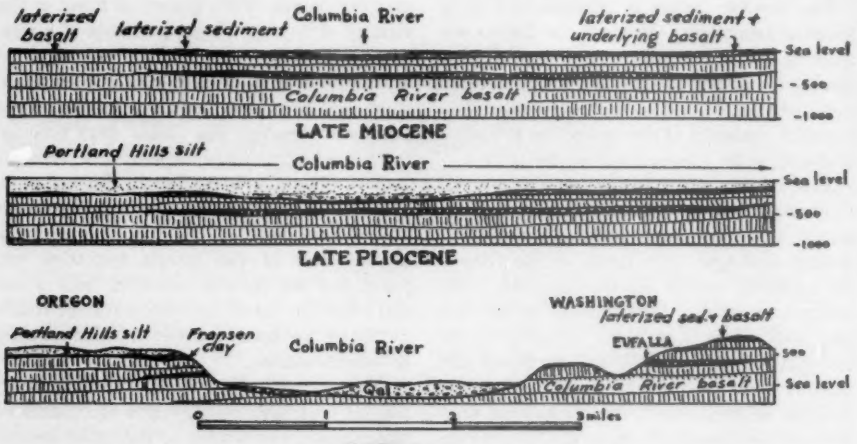


FIGURE 2.—SKETCHES SHOWING THREE STAGES IN THE DEVELOPMENT OF THE LOWER COLUMBIA RIVER VALLEY IN THE VICINITY OF MAYGER, OREGON, AND EUFALLA, WASHINGTON

instead breccias resembling volcanics poured out under water. This basalt breccia was found just west of the Astoria city reservoir and is thought to lie stratigraphically just below or to interfinger with the uppermost sandstone member described by Howe. Thus the post-basalt beds near Clifton are evidently younger than fossil-bearing Astoria strata which are intruded by Columbia River basalt feeders and overlain by extrusive basalt breccia.

The Astoria formation was evidently deposited in a localized subsiding basin as the part beneath the basalt horizon at Astoria attained a considerable thickness, whereas to the east the sediments are much thinner and the shale is missing. The uppermost Astoria sediments described near Clifton constitute the thickest section known to lie conformably above the basalt in the lower Columbia River Valley. No doubt the westernmost basaltic flows interfingered with shallow marine sediments as the basin became smaller in late Astoria time. Post-basalt sediments of this stage (Fig. 2) are found farther upriver in a shallow downwarp the Fransen clay pit 3 miles southeast of Mayger (Wilson and Treasher, 1938), in the cliff west of the mouth of Germany Creek (Fig. 2), and on U. S. Highway 30, 2.3 miles west of Rainier, Oregon where a bauxitic clay composed of decomposed pebbles lies on laterized basalt. Apparently the underlying basalt at the Fransen locality and westward was protected by a somewhat thicker section of upper Astoria sediments which were decomposed during laterization. These sediments were deposited during maximum submergence and their advanced state of decomposition indicates that the area lay low after the Astoria sea withdrew; otherwise erosion would have outstripped laterization.

Pliocene Formations

Introduction.—The Pliocene gravels, sands, and associated sediments of the lower Columbia River Valley are known as the Troutdale formation (Hodge, 1933). Those of the Clackamas River drainage have not been differentiated

from the Troutdale. The Pliocene sediments along the Molalla River are embraced in the Molalla formation (Nichols, 1944; Harper, 1946) and those of the North Santiam River are known as the Fern Ridge tuffs (Thayer, 1939). These fluviatile formations are essentially the same age (Plate 2). These sediments represent a filling of broad erosional valleys in Pliocene time as a result of submergence. Similar sediments may be present in other tributaries of the Willamette River (Baldwin and Howell, 1949). The fill is now found in several places in the Willamette Valley at elevations as great as 800 feet. As the Troutdale is the only one of the above formations which has been traced through the Coast Range, the Columbia River was probably in Pliocene time, as today, the only one of the above streams which emptied directly into the Pacific Ocean.

Contemporaneous with and interbedded in part with the Pliocene stream deposits are basic volcanic rocks of the Rhododendron, Boring "agglomerate," "Cascan," and Boring lava formations. The Rhododendron formation (Hodge, 1938) of the Bull Run and Sandy River drainage area and the Boring agglomerate (Treasher, 1942) of the Clackamas River are predominantly pyroclastic rocks which, however, were in part reworked by streams to form tuffaceous conglomerates or mud-flow breccias. Similar volcanic breccias occur in the Troutdale and Molalla formations and in the Fern Ridge tuffs. The Cascan formation (Hodge, 1938) is a thick series of andesitic and basaltic lavas whose pyroclastic phases, in large part water-worked, have been referred to as the Rhododendron. The Boring lava (Treasher, 1942) is a series of gray olivine basalts derived from a number of separate volcanoes. Although Treasher assigned the Boring lava to the Plio-Pleistocene, some of these lavas were noted by Williams (1916) to be interbedded with the Troutdale formation, and most, if not all, are restricted entirely to the Pliocene.

Troutdale formation.—The name Troutdale was introduced by Hodge (1933) for gravels and sands deposited by the ancestral Columbia River, which are well displayed near Troutdale, Oregon. Williams (1916) had earlier referred these gravels to the Satsop formation. The Troutdale formation was tentatively extended by Wilkinson, Lowry, and Baldwin (1946) to include a stratigraphically younger and much more widespread silt member for which the writers now propose the name Portland Hills silt. The silt may possibly be separated by a slight erosional break from the Troutdale formation proper, although both the silt and the Troutdale sands and gravels were deposited by the Columbia River as a result of a rise in base level accompanying submergence in Pliocene time. Williams and Hodge considered the Troutdale formation to be probably of Pleistocene age, but Chaney (1944) in his later study of the Troutdale flora assigned it to the lower Pliocene. Much earlier Washburne (1914, p. 27) had suggested that the silt which occurs on the Portland Hills was of Pliocene age on the basis of a fossil camel's tooth (*Auchenia*) found in it.

Although Pliocene sediments can be traced southward from the Troutdale formation (as mapped by Treasher, 1942, in the Portland area) into the coalescing Molalla formation, it is desirable to retain Hodge's definition of the Troutdale as, not only have other names (Molalla formation and Fern Ridge tuffs farther south) already been introduced, but also the composition of part of the Troutdale is different in that it contains quartzite pebbles. The Troutdale sands and gravels are restricted to the valley of the lower Columbia River, but the somewhat younger Portland Hills silt is not and covers much of the adjacent land.

DISTRIBUTION: Wilkinson, Lowry, and Baldwin (1946) pointed out that Troutdale gravels can be traced from a locality near Camas, Washington, which is across the Columbia River from Troutdale, northwesterly to a point north of Woodland, Washington where the formation is exposed best in the high cuts along U. S. Highway 99. There a massive interbed of basic volcanic breccia with a maximum thickness greater than 100 feet is present in the sands and gravels. Somewhat less indurated Troutdale gravel occurs farther down the Columbia River on the Washington side in a large gravel pit a hundred yards east of U. S. Highway 99 about 1 mile south of the Kalama River bridge.

Excellent exposures of quartzite-bearing gravels, tentatively correlated with the Troutdale, occur in a large gravel pit 1 mile north

FIGURE 3.—COLUMBIA RIVER GORGE AT CROWN POINT SHOWING THE UNCONFORMABLE RELATIONSHIP OF THE TROUTDALE FORMATION WITH THE UNDERLYING COLUMBIA RIVER BASALT

Uppermost part of Columbia River basalt is an erosional remnant of lateritic ferruginous bauxite

of Longview, Washington, and farther down the Columbia River on the Oregon side at the Quincy School at the mouth of Beaver Creek, 3 miles northeast of Clatskanie. Below the mouth of Beaver Creek, Troutdale sediments have been found only on the Washington side of the Columbia River. However, quartzite pebbles occur near Aldrich Point on the Oregon side in loose sediments which may be a part of the Troutdale formation. Farther downstream, horizontally but poorly bedded, indurated sandstone along the railroad west of Ivy Station may also be equivalent in age to the Troutdale formation. The sandstone appears to be of nonmarine, possibly estuarine, origin.

Troutdale gravels are exposed in several roadcuts on U. S. Highway 830 a little more than 2 miles east of Cathlamet, Washington, and also at the east city limits, where well-rounded quartzite pebbles and cobbles, some as much as 4 inches in diameter, occur along with basalt pebbles, cobbles, and boulders in a massive silty matrix. The Troutdale sediments in the Cathlamet area lie on the eroded surface of the Columbia River basalt.

Troutdale gravels can be traced up the Columbia River Gorge from the type locality at Troutdale nearly to Wyeth where rounded quartzite and other pebbles occur 2 miles south of the town on the flat crest of a spur at an elevation of 2700 feet (Allen, 1932, p. 56). Float of quartzite pebbles indicates its presence farther east in the upper reaches of Viento Creek (Williams, 1916).

Near Crown Point, the maximum elevation of the gravel is about 900 feet. As indicated by the log of the Ladd well (Piper, 1942, p. 132) in the Portland area, the base of sediments which probably belong to the Troutdale formation there is more than 1400 feet below sea level and 1700 feet below the surface. The total thickness of the Troutdale proper may have been nearly 2000 feet in the Portland synclinal basin.

LITHOLOGIC CHARACTER: The lower part of the sediments assigned to the Troutdale formation, as indicated by the Ladd well log, consists of fine micaceous sand, conglomerate, "marl", and some shale with iron-stained gravel and fossil plants. The hard rock on which they lie is presumed to be Columbia River basalt (Pl. 1).

In the vicinity of Troutdale, which is well up on the northeastern limb of the Portland syncline and higher in the section, the formation is made up largely of massive beds of conglomerate and tuffaceous sandstone. The sandstone contains a very high percentage of basic lava grains, and some of the conglomerate carries an appreciable percentage of well-rounded, light-colored quartzite pebbles and cobbles. These quartzite pebbles and cobbles which locally may be faceted (Holdredge, personal communication) appear to be a reliable guide to Troutdale deposits if the sediments are sufficiently indurated to stand in a nearly vertical bank. However, not all Troutdale gravel contains quartzite pebbles. The quartzite and other resistant pebbles in weathered Troutdale conglomerate are often characterized by a partial surface coating or staining of iron oxide.

CROWN POINT SECTION: At Crown Point about 6 miles due east of Troutdale on U. S. Highway 30, at a much higher elevation and also higher in the Troutdale section, the relationship of the upper part of the Troutdale formation to the Columbia River basalt, the period of laterization, and the Boring lava in this area is well shown. The accompanying sketch (Fig. 3) is similar to that of Williams (1916) with one important exception.

Williams showed the Troutdale gravels, which he referred to the Satsop formation, as disconformable on the Columbia River basalt, whereas the Troutdale sediments are separated from the basalt by an angular unconformity. The recent discovery of a remnant of ferruginous bauxite at the top of the basalt section near Crown Point on the Rooster Rock road indicates that both a period of laterization and a period of erosion preceded deposition of the Troutdale sediments. The angular discordance between the two formations is not as clearly shown. However, the structural break is indicated by the fact that the Columbia River basalt dips westward to disappear west of Corbett on the Columbia River at sea level, a little more than 2 miles to the west, whereas a flow of Boring lava in the Troutdale formation at Crown Point dips westward only slightly.

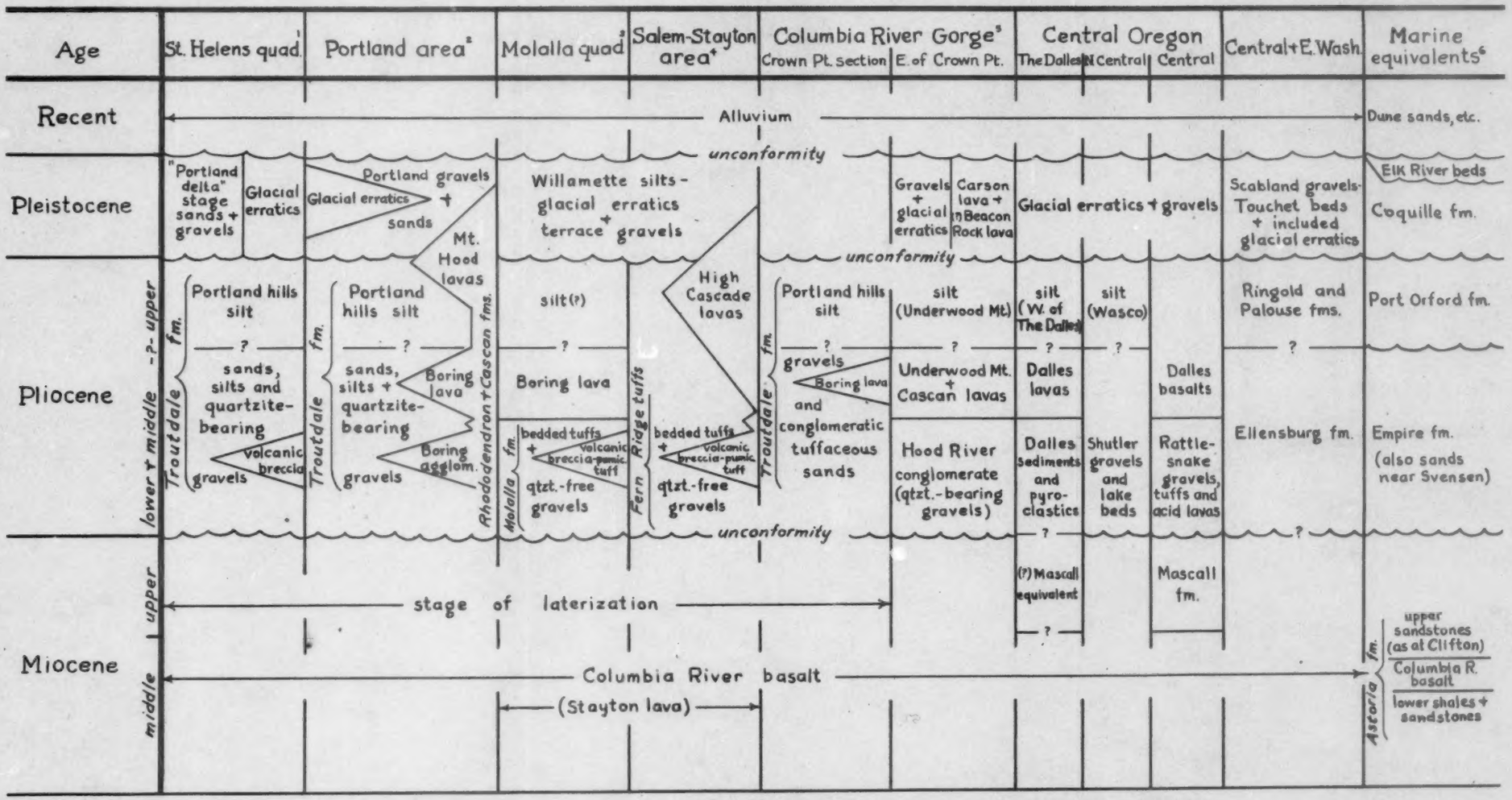
Folding of the Columbia River basalt prior to deposition of Troutdale sediments is also indicated by the thinness of the Troutdale at Crown Point as compared to that in the Port-

land area where downfolding of the basalt permitted the accumulation of a thick Troutdale section. Also, had not the Portland Hills anticlinal structure been largely developed in the Columbia River basalt before Troutdale gravels were deposited on the east slope, the waters carrying those gravels would have poured freely into the Tualatin Valley. However, no known Troutdale gravels are present there or on the crest of the Portland Hills. The angular discordance and erosional break between the Columbia River basalt and Troutdale formations at Crown Point may not be present in the bottom of the Portland synclinal basin nor in the Tualatin Valley at Hillsboro where a thick section of sediments presumably equivalent to the Troutdale but of a different texture occur above the basalt. The sediment in contact with the basalt in those two places is probably considerably older than the Troutdale sediments at Crown Point and may have been deposited as soon as downwarping of the laterized basalt began for, as already noted, the laterization is believed to have occurred while the basalt lay near sea level. The log of the Ray Maling well (Geological News Letter, 1937, p. 137) at Hillsboro in the Tualatin Valley suggests that about 20 feet of laterite has been preserved at the top of the basaltic section above which occurs 1400 feet of sediment. Perhaps all but the uppermost part is correlative with the Troutdale. However, it differs from the Troutdale in that it lacks gravels.

CROWN POINT BASALTIC INTERBED: At Crown Point and elsewhere, Williams (1916) noted that a lava flow was interbedded with the Pliocene gravel. Although originally described as a gray basic andesite about 30 feet thick, the flow is a porphyritic olivine basalt of Boring lava which for convenience may be referred to as the Crown Point lava member of the Troutdale formation. It dips 1° or 2° to the west and lies above sandstones and conglomerates of which some of the latter are characterized by steep torrential bedding and scour-and-fill structures.

SEDIMENTS ABOVE BASALT: Above the basalt flow is a 20-foot bed of decomposed conglomerate with basic lava pebbles and cobbles and a tuffaceous matrix similar to that in the unweathered conglomerate below the lava. The decomposed conglomerate is overlain, apparently conformably, by several feet of somewhat indurated massive sand. The sand contains scattered pebbles of quartzite and basalt. The basalt pebbles are decomposed like those in the underlying conglomerate but the quartzite pebbles are unaltered. The sand is incompletely exposed but it appears to grade upward into massive silt. The appearance of the quartzite pebbles in the massive sand does not need any special explanation, for they are common in some of the Troutdale gravels, but the change from the deposition of coarse tuffaceous sediments to a thick massive nontuffaceous silt is difficult to explain. Cessation of volcanic activity might well account for lack of volcanic material in the silt. Although a break in the depositional record is indicated here, it as yet does not appear to be of major importance. In the Portland Hills and in the St. Helens quadrangle, no great erosional interval is indicated; yet in several places the change from Troutdale gravel to massive silt occurs within a short vertical distance, without well-developed transitional phases. Instead, in these places and at Cathlamet, Washington, conglomeratic beds give way upward rather abruptly to massive silt.

PORTLAND HILLS SILT MEMBER: The writers propose the name Portland Hills silt for the uppermost member of the Troutdale formation. Although its exact relationship to the Troutdale proper is as yet unknown, the silt because of its much finer grain size and difference in composition is a distinct and mappable unit. At present it is considered to be the uppermost member of the Troutdale formation, but further work may show it to be of formational rank. The thick massive silt which overlies the Troutdale gravels at Crown Point is equivalent to that on the crest of the Portland Hills where in places it rests on laterized Columbia River basalt. Distribution of the silt, unlike that of the Troutdale proper, was not severely restricted by partially developed anticlinal structures in the Columbia River basalt for it forms a thick cover over much of the Portland Hills. Troutdale sands and gravels along Cornell Road at an elevation of 350 feet on the east slope of the hills give way upward to massive clayey silts which can be traced by discontinuous outcrops to the crest of the hills where the summit elevation is about 1125 feet. This



1- after Wilkinson, Lowry, and Baldwin (1946)
2- after Treasher (1942), Hodge (1938), and Lowry + Baldwin (this paper)
3- after Harper (1946)
4- after Thayer (1939), O'Neill (1939), and Allison and O'Neill (unpub.)
5- after Williams (1923), Hodge (1938), Allen (1932), Holdredge (1937), and Lowry + Baldwin (this paper)
6- after Baldwin (1945), Lowry and Baldwin (this paper), and others

CORRELATION CHART OF THE LATE CENOZOIC FORMATIONS OF THE LOWER COLUMBIA RIVER VALLEY

difference in elevation suggests that the silt may have a stratigraphic thickness of more than 700 feet in this area, but part of the difference probably can be explained by folding after deposition. Near the junction of Thompson and Cornell roads in sec. 30, T. 1 N., R. 1 E., a gritty lens in the silt shows stratification which is inclined several degrees to the east. Folding is thought to have taken place continuously while the sediments of the Troutdale proper and the Portland Hills silt were being deposited. Deposition of the silt member, like the rest of the Troutdale sediments, is attributed to a rise in the base level of the Columbia River during a rise in sea level in the Pliocene.

The silt on the Portland Hills was noted by Diller (1896) who first suggested its water-laid origin. He pointed out its similarity to eolian deposits, and Darton (1909) thought that it was wind-laid. Ruzek (1922) stated that the silt in Multnomah County, mapped as the Powell silt soil series, was at least in part of river origin and was laid down by the Columbia River when it flowed at a much higher elevation. Treasher (1942) stated that the heavy silt cover of the Portland area accumulated in late Pleistocene or early Recent times after the formation of the terraces and was of complex origin. He noted that part was of fluviatile origin and part was of eolian origin, although not as typically loesslike as that in the Palouse region of eastern Washington. Libbey, Lowry, and Mason (1945) pointed out that the presence of rounded quartzite and basalt pebbles in the silt on top of the Portland Hills indicates a water-laid origin, and they suggested that its present topographic distribution was largely the result of subsequent folding.

Scattered well-rounded quartzite pebbles which occur in the silt at an elevation of 1100 feet along the crest of the Portland Hills attest to its water-laid origin there. Similar pebbles have been found in the silt at elevations as much as 1500 feet farther west in Washington and Columbia counties where in places the silt disconformably and probably unconformably lies on extensive ferruginous bauxite deposits. A rolling topography has been developed on the silt in this area.

In most roadcuts where the Portland Hills silt has been exposed, it is a structureless, light-brown or yellowish-brown mass which in the upper part may be mottled with gray streaks. In deep cuts, the light-brown silt is seen to grade down into a reddish-brown, clayey silt. The silt is surprisingly uniform in grain size and in only a few places has a suggestion of stratification been noted.

Microscopic examination of more than 20 samples of silt collected between Crown Point on the east and Cathlamet on the west shows that it is quite uniform in composition. However, the silt contains very fine sand and pebbles in places. The component grains of most of the silt range in size from less than 0.05 mm. to more than 0.3 mm. Minerals identified in the silt include muscovite, biotite, feldspar including plagioclase, quartz, augite, hypersthene, hornblende, magnetite, tourmaline, garnet, apatite, and tremolite, as well as a very few grains of volcanic glass. Many of the samples contain opaline structures identified by Dr. Robert R. Shrock (personal communication) as sponge spicules. Diatoms are also present in the silt in places. The clay content of the upper part of the silt section is about 10 per cent, whereas the lower part, such as that at Sylvan in the Portland Hills, is sufficiently high in clay to permit its use in making brick and tile.

The Portland Hills silt is particularly well exposed in deep roadcuts along the Larch Mountain road near Crown Point. A 40-foot section of clayey silt exposed in one of the roadcuts on the east side of Ross Mountain shows faint stratification. Silt continues to the top of Ross Mountain at an elevation of 1200 feet where a well-rounded cobble is reported to have been found. The Troutdale formation there is essentially flat-lying and the contact between the conglomerate and the massive sand which grades up into the silt on U. S. Highway 30 near Crown Point is at an elevation of 900 feet, so the silt may be nearly 300 feet thick in this area.

In the Boring Hills, the silt is found on top of the hills at an elevation of at least 800 feet. Proof of water-laid deposits having covered the Boring lava to a height of at least 900 feet was furnished by Paul Howell (personal communication) who found scattered quartzite pebbles at an elevation of 900 feet in the NW¼ sec. 1, T. 2 S., R. 3 E.W.M., about 1 mile west of the

town of Boring, Oregon. Treasher's map of the Portland area (1942) shows that the core of this hill is Boring lava. The silt has also been found farther west along the Forsythe Road northeast of Oregon City and on top of Petes Mountain to the southwest at an elevation of 830 feet. Quartzite pebbles occur in the silt in a clay pit west of West Pioneer, Washington, which is north of Vancouver, Washington.

Along the Lone Oak Road north of Longview, Washington, quartzite pebbles occur in the silt in roadcuts at an elevation of nearly 900 feet. There it lies on laterized Columbia River basalt. The silt unconformably overlies bauxitic clay along U. S. Highway 30 and near Mayger, both west of Rainier, Oregon. A number of well-rounded quartzite pebbles were found in the silty soil 1½ miles northeast of Cathlamet, Washington, in the NW¼ sec. 6, T. 8 N., R. 5 W. at an elevation above 400 feet.

The occurrence of a camel's tooth in the silt on the Portland Hills prompted Washburne (1914) to suggest that it might be of Pliocene age. The sponge spicules and diatoms in the Portland Hills silt have not received much study. Although the silt cannot be assigned a definite age on the basis of its fossils, it is younger than the main period of Cascade volcanism. The silt contains only a very minor amount of material which can be attributed to contemporaneous volcanism, and most of the minerals in the silt were brought down the Columbia River from its upper basin.

The Portland Hills silt has been traced eastward to Underwood Mountain on the north side of the Columbia River near Hood River. There the silt extends to an elevation of 2200 feet on the slopes of the mountain which is a source of Pliocene lavas that in places lie on the Hood River conglomerate. The sequence is similar to that at Crown Point.

The Portland Hills silt member of the Troutdale is thought to be correlative with several formations farther east in the Columbia River basin. Similar silt has been found 1 mile west of Wasco, Oregon, at an elevation of about 1200 feet; along U. S. Highway 97, 2 miles northeast of its junction with Oregon Highway 50, at an elevation of 3175 feet; along the road to Clarno, 4 miles east of Antelope, Oregon, at an elevation of 3075 feet; and in the Blue Mountains near Kamela at an elevation of about 4325 feet. Samples from these localities contained many of the minerals found in the Portland Hills silt as well as similar opaline sponge spicules. Diatoms were noted in all except the Wasco sample. All these silts lie on the Columbia River basalt and all are thought to be of water-laid origin. The Wasco silt belongs to the Shutler formation which Hodge (1942) assigned to the Pleistocene yet also correlates with The Dalles formation now considered Pliocene. The difference in elevation of these silts is believed to be the result of differential uplift since their deposition, with the silt in the Blue Mountains having been uplifted the most.

The Portland Hills silt may also be correlative with the Ringold formation of south-central Washington and the Palouse formation of southeastern Washington and adjacent Idaho. The Ringold formation (Merriam and Buwalda, 1917) was originally assigned to the Pleistocene. However, Beck (personal communication) has referred fossils found in it to the upper Pliocene.

A sample of silt from the Ringold formation along U. S. Highway 395, 2 miles south of Lind, Washington, and a sample of Palouse silt (collected by Richard J. Anderson) from a roadcut 2 miles north of Moscow, Idaho, were found to be petrographically similar to the Portland Hills silt and to contain numerus opaline sponge spicules.

Culver (1937) pointed out that the Ringold formation extended as far east as McChesney Springs and considered much of the buff silt in the Palouse region its eastern equivalent. Although the beds in the Palouse region were long considered to be wind-laid, Treasher (1925) noted stratification in the silt and proposed the name Palouse formation. Rounded pebbles occur in the silt deposits and do not support a wind-laid origin. Although Bryan (1927) preferred the name Palouse soil, Scheid (1947) reintroduced the term Palouse formation but did not assign it an age. The Palouse formation appears to be correlative with the Ringold and, if so, is of late Pliocene or Pleistocene age.

The relative fineness of these deposits in the upper Columbia River basin indicates that erosion was relatively inactive during their deposition which is thought to have resulted from the rise in base level that caused deposition of the Portland Hills silt.

Although at least part of the Troutdale formation can be assigned on the basis of its stratigraphic relationships and its fossil floras, as well as the age of its correlatives, to the lower Pliocene, the lowermost part may be of late Miocene age and the youngest member, the Portland Hills silt, appears to be of late Pliocene or even early Pleistocene age.

STRUCTURE: The structure of the Troutdale formation proper and the uppermost Portland Hills silt member has been determined largely from the topographic distribution of the sediments rather than from attitudes. In most places in northwestern Oregon the attitude of the Troutdale formation is difficult to determine. Where attitudes have been obtained, most of the dips are low and some in part original. Steep original dips are present in torrential bedded gravel exposed near Crown Point.

The topographic distribution of the Troutdale sediments ranges from approximately 1400 feet below sea level in the Portland basin (probably 1200 for the correlative sediment in the Tualatin Valley basin) to 2700 feet above sea level on the south side of the Columbia River Gorge near Wyeth. Their occurrence below sea level in the Portland area is largely the result of downfolding of the Columbia River basalt while the sediments were being deposited, whereas the elevation of the quartzitic gravel in the gorge near Wyeth is largely the result of the vertical uplift of the Cascade Mountains at a later time. The topographic distribution of the Portland Hills silt may be explained mainly by deformation and uplift subsequent to its deposition. Although in places the silt is known to have been several hundred feet thick, its range in elevation from near sea level west of St. Helens to 1500 feet in Washington County is too great to be ascribed solely to original thickness. Also the lithological uniformity of the silt suggests that its original thickness in most places in the Portland and St. Helens areas was of the order of several hundred feet.

Molalla formation.—The Molalla formation (Nichols, 1944) consists of a series of fluviatile sediments which are widespread in the Molalla quadrangle and particularly well exposed along much of the Molalla River from a point below the town of Molalla upstream to a point above Trout Creek. As described by Harper (1946), the formation comprises sandstone, conglomerate, and silt which are highly tuffaceous for the most part and consequently light-colored. In many places, especially at or near terrace levels, the sediments are deeply weathered.

The tuffaceous sandstone and conglomerate occur in the lower part of the section. The beds show cut-and-fill structures and torrential bedding. The sandstone is friable and fossil leaves, which occur in it at several localities, are not well preserved.

A massive basic volcanic breccia which is interbedded with conglomerate and other sediments at the Molalla River bridge, east of the town of Molalla, resembles that in the Troutdale formation north of Woodland, Washington, and those within the Molalla formation itself along Milk Creek between Colton and Estacada. Treasher (personal communication) has noted its similarity to the Boring agglomerate. Very similar ones in the Fern Ridge tuffs occur farther south along Oregon Highway 214 west of Silver Creek Falls. Whereas the volcanic breccia in the Troutdale north of Woodland shows no definite signs of having been reworked, that in the Molalla and Fern Ridge tuffs does. Some of the chunks of basic lava show slight rounding, and pieces of carbonized wood are present.

The light-colored tuffs which occur in the upper part of the Molalla formation in the southeastern part of the quadrangle are nearly identical to and probably grade laterally into those in the Fern Ridge tuffs farther south. Massive pumiceous phases have been quarried in a few places for use as dimension stone.

A series of terraces, which roughly correlate with those farther north along the Clackamas River and those in the Portland area cut in the Troutdale formation, range in elevation from 250 to nearly 1000 feet. They cut the Molalla formation in the area around Molalla and also at much higher elevations in the narrower upper valley, as in the vicinity of Trout Creek. The Molalla clay deposit, which includes several million tons of high-alumina clay (Wilson and Treasher, 1938) underlies the gently sloping surface of the 650–700 foot terrace. Although a veneer of Pleistocene gravel laid down during the cutting of the terrace may be present, the bulk of the clay is believed to be an integral part of the Molalla formation.

The age of the Molalla formation is in dispute.

Piper (1942) correlated the sediments now known as the Molalla formation with the Troutdale. Fossil leaves from tuffaceous sandstone were assigned by Chaney (1944) to the lower Miocene. Sediments belonging to the Molalla formation can be traced from Molalla northeasterly to Estacada where Treasher (1942) included them in the Troutdale formation. The writers agree with Harper (1946) that the formation is Pliocene and correlate it with both the Troutdale and the Fern Ridge tuffs. The Molalla formation is not only similar lithologically but has similar stratigraphic relationships. The Boring lava of Pliocene age apparently overlies the Molalla formation conformably. The Molalla formation lies unconformably on Eocene basic lavas, breccias, and associated sediments. Unfortunately the Molalla formation has not been found in contact with the relatively restricted Columbia River basalt, but the basic volcanic pebbles in some of the Molalla conglomerate were derived in part from the Columbia River basalt upstream.

Pliocene sediments and the Boring agglomerate of the upper Clackamas River Valley.—If Hodge's definition of the Troutdale formation as those sediments laid down by the ancestral Columbia River is to be retained, then most of the sediment along the upper Clackamas River in the vicinity of Estacada which Treasher (1942) mapped as Troutdale perhaps should not be included. These sediments lack the quartzite pebbles typical of the Columbia River drainage which occur in the Troutdale formation only a short distance to the north and also farther down the Clackamas River in the Troutdale sediments in the Deep Creek drainage and on Clackamas Heights near Oregon City. However, the absence of quartzite pebbles is insufficient evidence to exclude them from the Troutdale formation.

The writers do not wish to propose a new name for the Pliocene gravels along the upper Clackamas River. The writers do believe that these sediments were being deposited by the ancestral Clackamas River while the ancestral Columbia and Molalla rivers, and farther south the North Santiam River, were depositing the sediments of the Troutdale and Molalla formations and the Fern Ridge tuffs, respectively.

A very few light-colored quartzite pebbles and cobbles were found south of Estacada along the Day Hill road on the 1000-foot terrace. However, they were not nearly as well-rounded as those in the Troutdale and were a different color and texture. Possibly these pebbles were brought by the ancestral Clackamas River from a quartzite terrain in central Oregon prior to the widespread extrusion of Pliocene lavas and the arching of the Cascade Range and the integration of the Deschutes River drainage.

The discovery of bauxitic material 2½ miles southeast of Estacada in secs. 3 and 4, T. 4 S., R. 4 E. was announced by the Oregon Department of Geology and Mineral Industries in 1948 (The Ore-Bin, vol. 10, no. 9, Sept., 1948). The bauxite has a higher content of silica and less iron than that in Washington and Columbia counties, and was formed not from Columbia River basalt but from the Pliocene sedimentary rocks with their associated volcanic rocks. The laterized sediments are reported to be overlain by Boring lava or its approximate equivalent. The laterization of these sediments occurred at a later time than that of the Columbia River basalt, for the Columbia River basalt in this area is unconformably overlain by the Pliocene sedimentary and associated volcanic rocks. The laterization appears to have preceded extrusion of the Boring lava and thus it would appear that it took place in a relatively short period of time.

Treasher (1942) introduced the name Boring agglomerate for a massive volcanic breccia which occurs in the vicinity of Estacada on the Clackamas River. The breccia is made up largely of angular fragments of basic andesite in a tuffaceous matrix, but it also contains, according to Treasher, rounded to subangular cobbles and boulders of Columbia River basalt and in places interbedded lava. He thought that the hot ash and associated debris, erupted from near-by volcanoes, picked up pebbles, cobbles, boulders, wood, and leaves when it fell to earth and after pouring over hillsides came to rest, partially filling canyons. Treasher believed that an unconformity exists between the Troutdale and Boring agglomerate; however, the Boring agglomerate appears to be made up of mudflows interbedded with the Pliocene gravels and associated sediments of the Clackamas River Valley. It is similar in appearance and composition to the breccia interbed in the Troutdale

north of Woodland, Washington and to those in the Molalla formation and the Fern Ridge tuffs.

If the Boring agglomerate is interbedded with the Pliocene sediments of the Clackamas River Valley, which are equivalent to the Troutdale formation, it is probably of lower or middle Pliocene age. It is equivalent in age to the Rhododendron formation of the Sandy River Drainage to the northeast which will be discussed later.

Fern Ridge tuffs.—The Pliocene sediments which filled the valley of the North Santiam River were first described by Thayer (1939) who called them the Fern Ridge tuffs, assigned them to the upper Miocene, and stated that they lie conformably on the Stayton lavas of Miocene age. Later O'Neill (1939) pointed out the unconformity between the Fern Ridge sediments and the basaltic lavas, herein included in the Columbia River basalt. The erosional break was established (Libbey, Lowry, and Mason, 1945) when partially rounded gibbsite nodules were found in Fern Ridge sediments. The presence of these nodules, formed by laterization of the Columbia River basalt, indicates that a long period of weathering and at least some erosion preceded the deposition of Fern Ridge sediments.

Most of the sediments constituting the Fern Ridge tuffs were laid down by the ancestral North Santiam River or fell in its valley where they are now found. Thayer's map of the North Santiam River and Salem Hills area (1939) and Allison's and O'Neill's map (unpublished) of the Stayton quadrangle show part of their occurrence. Erosional remnants of these deposits occur farther west than mapped, as in the Salem Hills, where rounded, much weathered pebbles have been found lying on laterized Columbia River basalt near the crest of the hills. Their occurrence at an elevation greater than 800 feet indicates that both the Salem and Eola hills were once covered by a fill through which the Willamette River was able to superimpose its course through these hills.

Eruptive materials are common in much of the Fern Ridge tuffs formation, as the name suggests, but most of it has been reworked by stream action although sorting and bedding are lacking or poor in many exposures. Pebbly conglomeratic beds are locally common in the lower part of the formation. In a roadcut southwest of the Union Hill school, the gravels show torrential cross-bedding with scour-and-fill structure. The gravel deposit is deeply weathered with many pebbles completely altered to clay. Water-sorted, buff to cream-colored tuffs are common. Many of the tuffs are decomposed, but grains of andesine feldspar in some of them are unweathered. Gray pumiceous tuff is exposed at several localities and it is only slightly weathered as compared with the associated or near-by reworked material.

Similar tuffs occur far up and on top of the south wall of the North Santiam River Valley at Mehama and also at similar and somewhat lower elevations on the north wall. Thayer included those on the north wall in his older Mehama formation despite their lithologic resemblance to the Fern Ridge tuffs. The tuffs well up on the south wall, which contain large chunks of silicified wood, were also included in the Mehama. However, the very similar tuff which higher up lies on top of Miocene basalt, rather than against it, was mapped as Fern Ridge tuffs by Thayer. All these tuffs are correlated with the Fern Ridge tuffs which fill the valley cut in Miocene basalt and the underlying Mehama formation.

North of the town of Mehama, conglomeratic beds of basic lava occur in the upper part of the Fern Ridge tuffs and are overlain by andesitic lavas of Pliocene or younger age at an elevation of about 2500 feet just west of House Mountain. These lavas are part of Thayer's High Cascade lavas, which are equivalent at least in part to the Cascan formation of Hodge (1938) farther north.

Thayer estimates that the thickness of the Fern Ridge tuffs is 1500 feet. Such a thickness is hard to explain unless the sediments filled a structural or erosional depression as postulated by the writers. The Fern Ridge tuffs show a slight westerly dip in most places where bedding can be seen. This dip may be in part original, but most is probably the result of the westward tilting of the western flank of the Cascade Mountains which accompanied uplift in late Pliocene or later time.

No vertebrate fossils or fossil leaves have been found in the Fern Ridge tuffs. The tuffs are correlative with the Troutdale formation of

Pliocene age, as both the Fern Ridge tuffs and the Troutdale formation lie on Columbia River basalt, and both are overlain by lava of Pliocene or younger age. The Fern Ridge tuffs can be traced northward into the Molalla River drainage where they have been referred to the correlative Molalla formation.

The topography of the Fern Ridge tuffs is characterized by several terrace levels which are believed to correspond roughly to those cut in the Molalla and Troutdale formations. However, the sediments far up the North Santiam River on the walls of the valley have steep slopes in many places, although there are some fairly wide, flat upland strips. Where the river-cut terraces have not been deeply eroded, the underlying sediment is deeply weathered. Some of the resultant clays have been described by Treasher (Wilson and Treasher, 1938). The sediments are not always easy to distinguish from Pleistocene alluvial terrace gravel such as that mapped by Thayer on the 500-foot terrace southwest of Stayton. There the Pleistocene gravels overlie, apparently disconformably, Fern Ridge tuff. The writers do believe that the bulk of the material making up the several terraces is part of the Fern Ridge tuffs with a relatively thin cover of Pleistocene alluvium.

Boring lava, Cascan, and Rhododendron formations.—Treasher (1942) introduced the term Boring lava for inflated gray olivine basalts in the Portland area which he assigned to the Plio-Pleistocene. He pointed out that these lavas in many places were derived from local sources of which Mt. Sylvania southwest of Portland and Green Mountain 5 miles north of Camas, Washington, are excellent examples. The age assigned to the Boring lavas by Treasher can be further restricted to include only the Pliocene as it is interbedded with the Troutdale at Crown Point and elsewhere lies below the Portland Hills silt. The Boring Hills, the type locality of the Boring lava, were completely covered by the Portland Hills silt and associated sediments, but much of the original blanket has been removed revealing their volcanic origin. Boring lava overlies the Molalla formation to the south and, as Treasher pointed out, the lava is essentially equivalent to at least part of the lavas of the Cascan (mnemonic for Cascade andesites) formation to the east. Hodge (1938) had previously included such volcanic sources in the Portland area as Mt. Tabor, Chamberlain Hill, Mt. Scott, and Kelly Butte in the Cascan formation. Mt Tabor is one of the few cinder cones of this age. Similar cinders were found at the base of the Portland Hills silt along Skyline Blvd., a few hundred yards north of its junction with Cornell Road. As at least some of the above lava buttes are predominantly of basaltic composition, Treasher may be justified in including them in his Boring lavas, leaving the name Cascan for less basic lavas.

Undoubtedly both basic and andesitic lavas were being erupted simultaneously in different parts of northwestern Oregon and southwestern Washington, and such volcanic cones as Mt. Hood and Mt. St. Helens were active in the Pliocene as well as in the Pleistocene. Volcanism in northwestern Oregon and southwestern Washington during the Pliocene was not restricted to any particular area. However, the most active region was the Cascade belt. Compressional folding was taking place at the time, and the greatest stress, as shown by the character of the folds developed in the Columbia River basalt, occurred in the Cascade Mountain belt. The porphyritic hypersthene andesite of Wind and Shellrock mountains in the Columbia River Gorge represents feeders to Cascan flows, as stated by Hodge (1938). The writers found that the massive columnar andesite of Shellrock Mountain passes up as a dike through Columbia River basalt into Cascan lava.

The Rhododendron formation is said by Sheets (1932) to be the pyroclastic phase of the Cascade volcanic activity in Pliocene time, and he called the interbedded and associated lavas the Cascan formation. As Sheets pointed out, the volcanic debris of the Rhododendron becomes less reworked as the volcanic sources are approached.

Lava of Boring type and of approximately the same age occurs in Underwood Mountain, as previously mentioned, and across the Columbia River on the Oregon side west of Hood River. Similar lava was derived from several volcanic cones near White Salmon, Washington. The olivine basalts now included in The Dalles formation are probably equivalent, at least in part, to the Boring lava.

Pliocene sediments of southwestern Washington.—In addition to the Pliocene fluviatile

sediments along the Washington side of the lower Columbia River below Portland which belong to the Troutdale formation, there are somewhat similar deposits farther north in the Cowlitz River and Chehalis River drainages. However, Pleistocene gravels are also present and their differentiation everywhere is not easy. Indurated gravels, in many places deeply weathered, and associated sediments are widespread in the Chehalis River Valley and are especially well exposed near Napavine and in roadcuts along U. S. Highway 99 south of Chehalis.

Many well-rounded, light-colored, mainly yellowish or light brown quartzite pebbles like those typical of the Troutdale formation occur 6 miles south of Chehalis in sec. 1, T. 12 N., R. 3 W. at an elevation of about 600 feet. None were found in the Pliocene (?) gravels to the east. A divide between the Chehalis and Columbia rivers is believed to have existed in Pliocene time until base level rose to a point corresponding to a present elevation of about 600 feet. The water, perhaps only floodwater, from the Columbia River apparently could enter the Chehalis Valley via the Cowlitz River Valley bringing in quartzite pebbles. As the Columbia River had already established its present shorter course to the sea, no diversion occurred.

The fluviatile sediments and associated highly vesicular lava in the Toutle River Valley east of Castle Rock in southwestern Washington resemble parts of the Troutdale and Molalla formations as well as the Fern Ridge tuffs and include the well known Castle Rock clay deposit described by Glover (1941) and later by Nichols (1943). Glover thought the deposit to be either of Cowlitz (upper Eocene) or of Pleistocene age. He stated that the nearly flat attitude of the deposit favored the Pleistocene age. Many of the Pliocene sedimentary deposits were first considered Pleistocene. Nichols favored a lower Oligocene age for the deposit. The clay deposit, or deposits, for there are several, are a part of erosional remnants of a Pliocene fill of the Toutle River Valley. The sediments of the Castle Rock clay deposits described by Glover in sec. 20, T. 10 N., R. 1 W. overlie Columbia River basalt of middle Miocene age, but elsewhere, such as farther upstream, they unconformably overlie basaltic lavas of probable upper Eocene age.

Quaternary Deposits

Early Pleistocene deposits.—The greater part of the early and middle Pleistocene was a time of active erosion as the streams kept pace with regional uplift. There were still-stands of the rivers which no doubt coincided with some of the high coastal terraces. During such pauses gravels were left on terraces cut in Pliocene and older Tertiary strata. Allison (1936) has described some of these older Pleistocene terraces in the Willamette Valley and suggested possible correlation with the glacial stages.

These relatively thin deposits are generally more weathered than the later Portland gravels.

Portland gravels and included glacial erratic boulders.—The Portland gravels, sometimes referred to as the "Portland delta gravels", occupy a prominent place along the Columbia River between the Sandy River on the east and the Willamette River on the west. The gravels were deposited mainly by the Columbia River during the Pleistocene. Correlative gravels occur to the south in the Willamette Valley. Finer-grained deposits, predominantly sands, occur down the Columbia River from Portland and are well exposed in places in the Cowlitz River Valley below an elevation of 350 feet.

The surface of the Portland gravels, a considerably dissected terrace, has a maximum elevation of approximately 365 feet above sea level, and beds in the lower Sandy River Valley, which may represent the uppermost part of the series, are about 400 feet in elevation. The base of the gravels is not known with certainty, but as interpreted from well logs, it is in places 200 feet below present sea level (Piper 1942, p. 34, Pl. 7). Thus the maximum thickness may be nearly 600 feet.

In the vicinity of Portland, the sediments are predominantly sand and fine gravel. Coarse gravel with cobbles and even large boulders is common in east Portland, whereas the sediments are finer-grained in north Portland near the peninsula at the confluence of the Willamette and Columbia rivers. The pebbles are generally well-rounded and mostly of basaltic composition; Cascan- and Boring-type lavas are common as well as those of Columbia River basalt. Most of the extremely large boulders are Boring lava, which need not have traveled far, but large blocks of Troutdale conglomerate

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are also present. Boulders of plutonic and metamorphic rocks occur but are relatively uncommon; most of these are glacial ice-rafted erratics. Quartzite pebbles, no doubt largely reworked from Troutdale sediments, are included. Deep cuts in north Portland show foreset beds of sand and fine gravel built up layer by layer. The foreset bedding in most places is inclined to the south or southwest.

Excavation for a pipeline paralleling Clay Street in southwest Portland exposed about 200 feet of light-brown to medium-gray, poorly stratified, medium-grained, micaceous sand. The presence of quartz grains and mica points to reworking of older sediments by the Willamette River during the stage of alluviation. Nothing coarser than sand was observed within this section to indicate flooding of unusual magnitude during deposition.

GLACIAL ERRATICS: The plutonic and metamorphic boulders in the Portland gravels are mainly granitic erratics which could have come only from east of the Cascades. Allison (1935) noted the occurrence of foreign cobbles and boulders as much as 3 feet in diameter at depths as great as 30 feet in Portland gravels. Treasher (1942) noted the occurrence of occasional foreign or "erratic" boulders in the formation. Paul Howell (personal communication) noted several large unrounded boulders in gravel pits near Troutdale at an elevation of approximately 225 feet. Several angular erratic boulders have been found in other pits just east of Portland at an elevation of about 300 feet. Allison concluded that the inclusion of fragments of granite and quartzite in a matrix of silt at depths of 6 feet or more indicated ice-rafting. He considered the erratics in the Portland gravels to have been deposited by the same agency and at approximately the same time as the other glacial erratics in the Willamette and Tualatin valleys. Allison stated that the erratics were ice-rafted from their source in the upper Columbia River basin down the Columbia River into the Willamette Valley in early Wisconsin time (Spokane stage). The angularity of some of the erratics in the Portland gravels makes their transportation in icebergs imperative.

DEPOSITIONAL HISTORY: Allison stated that the erratics in the Portland gravels are equivalent in age to those elsewhere in the Willamette Valley, which as he points out are associated with the Willamette silts—a still-water facies of the Portland gravel stage. Glacial erratics of the same type occur in the St. Helens quadrangle (Wilkinson *et al.*, 1946). Perhaps many of the erratics were deposited late in the alluvial cycle as indicated by the upper limit of some, but some are apparently related to later flooding. However, the wide distribution of the erratics, the topographic range of those incorporated in alluvial deposits, and the common maximum elevation of the Portland gravels and Willamette silts, as well as their included erratics, indicate that the Portland gravels, Willamette silts, and at least part of the erratics were deposited in a body of water as it rose to a point represented by a present elevation of about 400 feet. The evidence points to a rise in base level governed by a eustatic rise in sea level. Based on Allison's assignment of the erratics to the Spokane (Wisconsin) stage of glaciation of eastern Washington, this apparent rise in base level occurred in late Pleistocene time, and thus the Portland gravels would belong to the same stage. The Spokane stage of glaciation of eastern Washington is older than the last or late Wisconsin stage and presumably early Wisconsin. It may prove to be somewhat older.

If the bulk of the Portland gravels represents one stage of alluviation by the Columbia River, as believed, their apparent thickness of more than 500 feet may be explained largely by a eustatic rise in sea level following a period of glaciation during which time downcutting by streams amounted to several hundred feet. The withdrawal of water onto the land in the form of snow and ice in early Wisconsin (?) time appears to have been by far the greatest of any in the Pleistocene. Possibly uplift in the Portland area was somewhat more rapid earlier in the Pleistocene and tended to offset the eustatic rises in sea level so that stages of alluviation then were not as extensive. However, uplift since deposition of the Portland gravels has been 400 feet (assuming that the amount of water tied up on land now is approximately the same as that at the height of the Wisconsin eustatic rise of sea level), whereas the uplift since the deposition of the late (?) Pliocene Portland Hills silt may not be much more than 1100 feet in the Portland area. Thus it would appear that either the lowering

of sea level accompanying early Wisconsin (?) glaciation was greater than previous ones or that its duration was longer so that the Columbia River was able to cut a deeper channel in which a thick alluvial fill could accumulate.

The apparent much greater downcutting of the lower Columbia River Valley in early Wisconsin (?) time prior to deposition of the Portland gravels agrees with Fisk's findings in the lower Mississippi Valley (Fisk, 1939). There five oscillations in base level of streams, which are indicated by five alternations between erosion and deposition, are correlated with the cyclic eustatic changes in sea level accompanying Pleistocene glaciation and deglaciation stages. As shown by one of his figures, the downcutting during early Wisconsin time was greatest, and the thickness of the fill occupying the resultant valley is more than 275 feet.

The events during the deposition and later dissection of the Portland gravels correspond closely with the Pleistocene history of the Oregon Coast as outlined as follows by Baldwin (1945): (1) General uplift and erosion of the Pliocene deposits were accompanied by the formation of marine terraces. Uplift continued (relative to sea level) to a point estimated to be about 250 feet below present sea level and then (2) a stage of submergence occurred which was probably caused by a eustatic rise of sea level accompanying deglaciation. The deposition of the estuarine sediments of the Coquille formation of late Pleistocene age accompanied submergence. (3) General recession of the sea following maximum Coquille-stage filling was accompanied by the formation of lower marine terraces of which the one capped by the Elk River beds of late Pleistocene age (elevation because of warping ranges from sea level to 225 feet) is the most prominent along the coast. Most of the present bays then were excavated considerably below present sea level. This stage of downcutting probably corresponded to late Wisconsin glaciation. (4) A relatively recent rise in sea level which accompanied the last stage of deglaciation has drowned the river mouths for many miles inland.

In the Portland region the history appears to have been similar. The Columbia River cut its pre-Portland gravel valley below present sea level into the Troutdale and other formations. Downcutting below present sea level by the Willamette River apparently extended only to the Willamette Falls at Oregon City. The deposition of the Portland gravels apparently occurred at the same time as that of the beds of the Coquille formation. As sea level lowered, several river terraces were formed upon the Portland gravel fill while marine terraces were being developed on the coast. Floodplain deposits younger than the bulk of the Portland gravels, such as the sands along the south side of the Columbia River near Rose City and in places along Columbia Blvd., may have veneered the gravels during this stage of terracing. These sands are less than 100 feet in elevation. The gravels may have been reworked, in some cases leaving lag gravels as a veneer.

Clackamas gravels, and Willamette terrace deposits.—Treasher (1942) recognized gravels in the Clackamas River Valley similar to the Portland gravels in stratigraphic position and elevation. He likewise discussed terrace deposits in the Willamette Valley of equivalent age. Gravels of this stage along the Clackamas River are as much as 400 feet in elevation near Estacada and form a thin veneer over Troutdale sediments.

Willamette silts and equivalent silt in the Tualatin Valley and associated glacial erratics.—Whereas the more vigorous Columbia River and its Cascade tributaries were depositing coarse sediments in the Portland region in early Wisconsin (?) time, the broad Willamette and Tualatin valleys were at times covered by standing water. West of the gravel deposits that are found on the east side of the Willamette Valley and in lesser amounts on the east side of the Tualatin Valley and near the mouth of the Tualatin River lies a broad dissected plain that averages between 180 and 200 feet in elevation. The plain is now youthfully dissected and flanked by a lower floodplain paralleling the streams. Silts immediately underlying this plain in the Willamette Valley have been referred to as the Willamette silts by Allison (Albany quadrangle, unpublished). In both valleys, especially the Tualatin, the silt rests on the thicker Portland Hills silt.

Allison (1935) noted that the valley-fill is not restricted to the floor of the Willamette Valley but extends as a mantle of ash-gray to buff silt with erratic fragments up on the walls in decreasing thickness to an elevation of about 400

feet above sea level. A similar mantle was noted in the Tualatin Valley near Helvetia. This may be the result of late flooding after deposition of the thick silt in the Willamette Valley.

Silt of the general valley-fill is exposed in a clay pit near the Yamhill River near McMinnville at an elevation of about 200 feet. The silty clay is roughly banded, more from color variation than from textural changes, into beds about 10 inches thick. During the past few years several dozen angular erratics as much as 14 inches in diameter have been uncovered and discarded by the operation.

There is no good evidence that the general sedimentary filling in either the Willamette or Tualatin valleys exceeded 200 feet. However, the fill is higher near the eastern margin of the Willamette Valley where the Cascade tributaries of the Willamette River filled their valleys to greater heights. As the main fill is thought to have reached only about 200 feet, the remaining part of the 400-foot level coinciding with the top of the Portland gravels was ponded water and the probable result of an eustatic rise in sea level and not temporary flooding.

Allison has assigned the Willamette silts and associated erratics to the Spokane stage of Wisconsin glaciation of eastern Washington. As already noted, the Willamette silts are correlative with the Portland gravels.

Late Pleistocene flooding and glacial erratics.—The Portland gravels and related Willamette silts do not show evidence of deposition during unusually large flood stages. There are sediments in the Portland area though that are torrentially bedded and which appear to be younger than the Portland gravels. Poorly sorted gravels in gravel pits near Rocky Butte, coarse and poorly sorted deposits along the Columbia River near Scappoose and Deer Island, and the gravel deposits at the west end of Oswego Lake appear to be younger. There is evidence of flooding (Allison, 1932) across the Portland gravels in east Portland where channels are cut in the gravels at an elevation of 200 feet around Rocky Butte. Flood water from the Columbia River appears to have poured through the Oswego Lake Gap as proposed by Allison (1935). Abundant limonite pebbles similar to the limonite found to the east in Iron Mountain on the north side of Oswego Lake and a few granitic erratics occur in the Tualatin gravels and indicate that the depositing waters moved westward through the Oswego Lake Gap. The Tualatin gravels were mapped by Treasher who correlated them with the Portland gravels. However, the Tualatin gravels in the area west of Oswego Lake appear to be younger than the silts and sands elsewhere in the valley which form the terraces along the Tualatin River and which do correlate with the Portland gravels. Pardee (1942) presents evidence of the release of a large quantity of flood water in the upper Columbia River drainage. Flooding in the Portland area probably coincides with this stage of flooding.

Evidence of a large flood of water passing through the Columbia River gorge has been noted by S. C. Sargent of the Corps of Engineers, U. S. Army, to whom the writers are indebted for the following conclusions:

> Flood water reaching an elevation of approximately 1,000 feet in the gorge between Wallula Gateway and The Dalles swept away talus deposits and veneered existing beds of gravel and sand with poorly sorted material. In addition spill-over channels with related deltaic deposits were formed. West of the gorge the water was not confined to a narrow gorge and spread out through such channels as the Lacamas Creek Valley to the north and spill-over channels across the Portland gravels to the south. The time of flooding was quite late in the Pleistocene as the river had cut down to approximately its present level prior to flooding and flood deposits have been very little eroded. The flood did not deposit normally bedded alluvium in the valley but instead modified an already incised surface, part of which was underlain by an earlier gravel fill coextensive with the Portland gravels.

This flood is in many ways similar to the Spokane flood described by Bretz (1925) but differs both in time and effects. Some scouring occurred but a relatively small amount of sediment was deposited by the flood. It quite adequately explains why the erratics east of the gorge are much higher than those to the west.

Although the writers conclude that many of the erratics in the lower Columbia River Valley were deposited with the Portland gravels and the Willamette silts, many may have been flood borne still later in the Pleistocene.

Gravels in the Columbia River gorge and upstream.—Gravel deposits resembling the Portland gravels lithologically occur in the Columbia River Gorge at Bonneville Dam, Wind River Valley and elsewhere. Holdredge (1937) reported fresh gravel deposits with included

granitic erratic boulders in the vicinity of Bonneville Dam. The Wind River deposit may be a post-Portland gravel flood deposit as the foreset beds are inclined upstream away from the Columbia River.

Relatively fresh gravels whose history of deposition may be comparable to that of the Portland region were noted by Piper (1932, p. 132) in the vicinity of the Dalles. The top of the gravel fill which is thought to correspond to the top of the Portland gravel fill is perhaps higher than normal aggradation would allow. Piper (1932, p. 132) stated that the fill reached to or above the 600 foot wind gap between the Columbia River and Fifteenmile Creek. The base of the gravels, which is about 150 feet in elevation, is much higher than that in the Portland region. The present Columbia River is obviously far from grade and the rapids at The Dalles may be in large part attributed to recent regional uplift of the area east of the gorge. Quite recent faulting at The Dalles and uplift of the eastern side was noted by Piper (1932, p. 139). Most of the faults as mapped show that the eastern block was upthrown, a fact which probably contributes to the presence of the rapids at The Dalles. Rapids near Umatilla and at the Wallula Gateway point to additional uplift upstream. It is altogether probable that warping and movement along the faults has occurred throughout the late Pleistocene.

Scabland gravels—Touchet beds.—The Scabland gravels and Touchet deposits of central and eastern Washington are tentatively correlated with the Portland gravels. According to Lupher (1944, p. 1436), fresh gravels and sands of the Scablands interfinger with the Touchet beds of fluvio-lacustrine origin. Both deposits extend from about river level at approximately 350 feet up to 1100 feet in elevation in the Pasco-Walla Walla, Washington region, although the thickest deposits of Touchet sediments are on the valley floors and only a thin veneer of sediments and some ice-rafted erratic boulders are found near the maximum elevation. The higher veneer of silts and higher erratics may be attributed to the late stage of flooding recognized by Sargent. The thicker deposits in the Pasco Basin are discussed below. The gravels are confined to the Scabland channels whereas the Touchet silts and associated sands and gravels cover the broader basins along the Columbia and Snake River drainages. Thus it would appear that they were deposited in a fluctuating body of shallow water in which fluvial deposits interfingered with lacustrine beds. Ice-rafted glacial erratics are common in the Touchet beds.

Lupher (1944) has indicated that the Scabland gravels and Touchet beds may be treated as contemporaneous. They contain erratics similar to those found in the Portland gravels. The writers suggest that the Scabland gravels and the bulk of the Touchet beds are not only contemporaneous with the Portland gravels but also were deposited as a result of the same submergence as that which affected the Portland region. The thickness, cyclic nature, and probable fluvio-lacustrine origin of the Touchet beds were noted by Lupher (1944, p. 1446). Deposition appears to have taken place in a gradually rising body of water unlike that which would result from unpredictable fluctuation accompanying ice jams or dams or by flooding (Flint, 1938). Thin deposits of silt, gravel and erratics were no doubt deposited by a late flood, but flooding occurred long after the Portland gravel stage.

Glacial erratics east of the Columbia River gorge.—The higher elevation of the erratics east of the gorge when compared with those to the west has puzzled many writers. Many large erratics have been found at approximately 1000 to 1100 feet altitude between The Dalles and Wallula Gateway and at even higher altitudes beyond the Gateway. Hodge (1931) reports erratics at even higher elevations farther south. It is probable that the erratics were contributed by several stages of deposition.

It has been suggested by the writers that many erratics were deposited during deposition of the Portland gravel and Touchet beds. Evidence of a large flood, first suggested by Bretz but substantiated by Pardee (1942) and Sargent (personal communication) no doubt furnishes the source of many erratics, particularly those near the 1000-foot elevation in the area near The Dalles and eastward which coincides with the top of the flood waters. Still others may have been contributed by older formations through reworking. Sargent has suggested that the flood water was restricted by the narrows at both the Wallula Gateway and west of The Dalles so

that it was backed to a higher elevation upstream from the Gateway, perhaps to the 1600-foot elevation mentioned by earlier writers, whereas below the Gateway it reached approximately 1,000 feet and west of the gorge it spread out rising little above 400 feet in the Portland area.

Development of the Lower Columbia River Drainage

The Columbia River is an antecedent stream. It established what is essentially its present course through the Cascade Mountains and the Coast Range after the extrusion of the Columbia River basaltic lavas in middle Miocene time. Subsequent alluviation of its valley in the Pliocene and the uplift of the Cascade Mountains and Coast Range across its course in later Pliocene and Pleistocene time have not significantly altered that course.

The original course through the Cascade Mountains was developed on Columbia River basalt. The much greater thickness of the basalt in the Columbia River Gorge than in the areas adjacent to the north and south indicates a preexisting valley through the ancestral Cascade Range, which formed the eastern shore of the Oligocene sea. The Eagle Creek conglomerate, which underlies the Columbia River basalt in the gorge area, is tentatively correlated with the Oligocene marine sediments farther west, and its deposition, possibly by the ancestral Columbia River, may have resulted from a rise in base level accompanying the invasion of the Oligocene sea.

Farther downstream, as between St. Helens area and Longview, the Columbia River evidently established itself at the margin of the Columbia River basalt where the lavas flowed against the higher terrain of Eocene basic lavas and associated sediments to the east and north. It established its course through the Coast Range on the Columbia River basalt.

The gradient of the Columbia River at this time was low and its erosive power slight, as indicated by the extensive development of laterized basalt on the Columbia River basalt.

In Pliocene time, while folding was taking place, alluviation first occurred where downfolds were being developed along the course of the Columbia River as in the Portland and St. Helens areas, where the oldest Troutdale deposits are found. As sea level rose in the Pliocene, alluviation of its entire course took place. Tributary streams draining northward into the Columbia River included the ancestral Clackamas, Molalla, and North Santiam rivers. When sea level was at its maximum height in late Pliocene(?) time, alluviation of the lower Columbia River and Willamette valleys was at its peak. The widespread deposition of the Portland Hills silt in the Tualatin Valley and the equivalent fill in the Willamette Valley led to the establishment of the present Willamette and Tualatin rivers. The present courses of the Kalama and Coweman rivers of southwestern Washington differ somewhat from their pre-Troutdale courses.

The course of the Columbia River through the Cascade Mountains was not affected much by the deposition of the Troutdale fill, for the river occupied a relatively narrow channel from which it could deviate only slightly. Although the river was antecedent to the uplift of the Cascade Range, the present course is in a strict sense superimposed from the Troutdale fill, whose upper part consisted of the Portland Hills silt. In places the courses naturally coincided.

Hodge (1938) however, believes that the course of the ancestral Columbia River lay south of the present channel and that lavas from Mt. Hood dammed it, producing ponded water which finally overflowed at the site of the present gorge. Hodge argues that, had the Columbia River established its present course across the Cascade Mountains prior to the extrusion of Cascade lavas, the lavas from such Cascade feeders as Wind and Shellrock mountains would have flowed into the gorge. The writers agree with Hodge that Wind and Shellrock mountains were feeders to Cascade lavas exposed higher in the gorge walls, but the present gorge had not yet been cut. The gorge itself is the result of incision by the Columbia River resulting from the uplift of the Cascade Range in post-Troutdale time. The presence of typical Troutdale sediments with quartzite pebbles within the gorge itself high on the canyon walls indicates that the Columbia River flowed along its present course in Troutdale time. The Troutdale sediments in the north-south valleys described by Hodge were, according to him, derived by streams whose sources lay north of the Columbia River, but no terrain is known there capable of giving rise to the quartzite pebbles in the Troutdale. The Troutdale gravels north and east of

Washougal, Washington, can be explained much better as a back-filling by the aggrading Columbia River rather than as a Pringle channel fill. The southerly dip of the gravels on the south side of the gorge may also be the result of back-filling. Unless the Columbia River had been superimposed from a Troutdale fill, it is unlikely that its course would have been superimposed across the no longer present volcano or volcanoes fed by magma coming up along the present sites of Wind and Shellrock mountains between which the river now flows.

The present height of the Cascade Mountains, although in part the result of the accumulation of Pliocene and Pleistocene lavas, is dominantly the result of regional uplift. The amount of uplift can be estimated from the maximum elevation in the gorge area of Troutdale gravels, which are believed to have been deposited near sea level. South of Wyeth, the gravels have been found at an elevation of 2700 feet. The 2200-foot elevation of the silt on Underwood Mountain indicates that the uplift there has been of similar magnitude. Hence the maximum uplift has been at least 2700 feet and perhaps somewhat more. The uplift has been differential, greatest near the axis of the Cascade Range and much less to the east and west. As uplift progressed, the streams continued to cut down their channels. The Columbia River began to cut its present youthful gorge down through Pliocene sediments, lavas and their feeders into Miocene and older lavas and sediments.

Uplift of the Coast Range has probably been somewhat less than that of the Cascades. However, laterized Columbia River basalt occurs at an elevation of approximately 1150 feet near Keyser, 5 miles due east of Cathlamet. The basalt attains its maximum elevation of more than 3000 feet on Nicolai Mountain south of the river. The course of the Columbia River across a limb of an anticline indicates that this part of its course is also antecedent.

Although uplift was continuous in the lower Columbia River basin, eustatic rises of sea level in the Pleistocene resulted in the formation of a series of terraces. However, in the lower Columbia River basin no well-preserved alluvial or depositional terraces with an average elevation much greater than 400 feet are known. The well-preserved 800-foot terrace along the Columbia River north of Washougal, Washington, the 550–600 foot terrace east of Woodland, Washington, and corresponding levels elsewhere are mainly the result of the planation in Pleistocene time of Troutdale and correlative sediments. They may have a veneer of Pleistocene gravel but it would be difficult to distinguish. Very little or no reworked Cascan or other Pliocene lavas have been recognized in them, although Columbia River basalt is a common constituent.

Downcutting by the Columbia River and tributaries during the early Wisconsin (?) glacial stage preceded deposition of the Portland gravels. The Columbia River was able to keep its channel nearly filled with sand and gravel as sea level rose during subsequent deglaciation to a point which is now nearly 400 feet above present base level as a result of subsequent uplift. The Cowlitz River likewise filled its channel with sand during the same eustatic rise. The Willamette River could not fill its broad valley as fast as sea level rose, so that the valley floor was covered by a large lake on which icebergs from the Columbia River floated and in which a blanket of silt was deposited. However, the Cascade tributaries of the Willamette River did fill their upper valleys to form terraces whose height corresponds to that of the Portland gravels.

A second stage of downcutting, accompanying and accelerated by uplift in late Wisconsin time, has amounted to approximately 400 feet in the Portland area. Uplift in the Columbia River Gorge and farther east has been even greater.

Although the upper courses of the streams are being eroded, the lower courses are drowned as a result of subsequent rise in base level accompanying deglaciation.

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METAMORPHIC FACIES IN THE WISSAHICKON SCHIST NEAR PHILADELPHIA, PENNSYLVANIA

BY DOROTHY WYCKOFF

ABSTRACT

The mica schists and mica gneisses of the Wissahickon formation in the Philadelphia region are described in terms of metamorphic facies. Representatives of the amphibolite facies (sillimanite-almandine and staurolite-kyanite subfacies) predominate, but show incipient alteration to minerals characteristic of the epidote-albite amphibolite facies. The status of rocks containing the assemblages kyanite-orthoclase, sillimanite-muscovite, and kyanite-almandine is also discussed.

Metamorphism of the highest grade is found in the southwestern end of the schist belt; but evidence is presented to show that the most intense metamorphism took place, not at the time of highest temperatures, but during a succeeding period of declining temperatures, when mineral changes were facilitated by copious hydrothermal solutions and strong regional deformation. The decipherable history is therefore largely one of retrograde metamorphism; many of the higher grade rocks have been converted by granitization to microcline gneiss ("granodiorite"), especially in the southern part of the schist belt, while farther to the north and west, incipient chloritization has been favored by late crushing. The period of metamorphism is tentatively dated as Paleozoic.

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Introduction

Two of the most important concepts underlying modern interpretations of metamorphic terranes are the concept of *zones* or *grades* of regional metamorphism and the concept of metamorphic *facies*. Metamorphic zoning was first demonstrated in the mapping of the Scottish Highlands (Barrow, 1893; 1912), and has since been applied to other regions. Metamorphic facies were first discussed by Eskola (1915; 1920) and the idea has been elaborated by many other workers. Attempts have also been made to correlate these two ways of grouping metamorphic rocks: a clear statement is given by Turner (1948, p. 76).

The two concepts are not mutually contradictory, but they do differ somewhat in emphasis. The concept of metamorphic zones or grades emphasizes the effect of varying physical conditions upon material of constant chemical composition. The zonal boundaries—delimited by mapping the distribution of certain "index" minerals such as biotite, almandine, staurolite, kyanite, and sillimanite in pelitic sediments—are assumed to correspond to lines of equal temperature and have been designated isograds (Tilley, 1924) to indicate that they are lines along which metamorphism has reached the same grade. Thus, on the basic assumption that the rocks selected as a standard have the same initial composition, schists containing kyanite or staurolite are considered to have reached a higher grade of metamorphism than those which contain only almandine or biotite as an index mineral.

The concept of metamorphic facies, on the other hand, provides for consideration of the effect of varying chemical composition as well as varying physical conditions. The different facies are distinguished as representing different sets of physical conditions: in regional mapping inter-facies boundaries would correspond to isograds. Within a single facies, however, differences in chemical composition may give rise to rocks of widely differing mineralogical make-up. For example, the amphibolite facies may include rocks characterized by biotite, almandine, staurolite, or kyanite—all formed simultaneously under similar conditions of temperature, pressure and shearing stress. Schists containing staurolite or kyanite are not therefore interpreted as rocks of "higher grade", but merely as rocks somewhat richer in Al_2O_3 and poorer in K_2O than schists which contain only micas.

This point is illustrated by the two facies diagrams (Figs. 1, 2) representing possible mineral assemblages within the amphibolite facies. I have adopted Turner's suggested names (Turner, 1948, p. 81–87): staurolite-kyanite subfacies (Fig. 1) and sillimanite-almandine subfacies (Fig. 2). In these systems $A = Al_2O_3 - (CaO + Na_2O + K_2O)$; $F = FeO + MnO + MgO$; $K = K_2O$. It is assumed that CaO and Na_2O are combined with Al_2O_3 in plagioclase, which may occur as an additional phase; and if SiO_2 is present in excess, quartz also appears, making 5 main phases possible in each assemblage.

A rock with the initial composition represented by point R would, under the physical conditions characteristic of the staurolite-kyanite subfacies (Fig. 1), be a muscovite-biotite schist. Such a rock cannot by "progressive metamorphism" develop in succession the index minerals almandine, staurolite, kyanite, unless at the same time its chemical composition is progressively altered along some such course as is indicated by points S, T, U: either Al_2O_3 must be added or K_2O, FeO, and MgO must be lost as metamorphism proceeds.

For comparison, a rock of the same composition, R′, is shown on Figure 2, which represents the assemblages stable under the physical conditions of the sillimanite-almandine subfacies. This rock would be a schist or gneiss containing

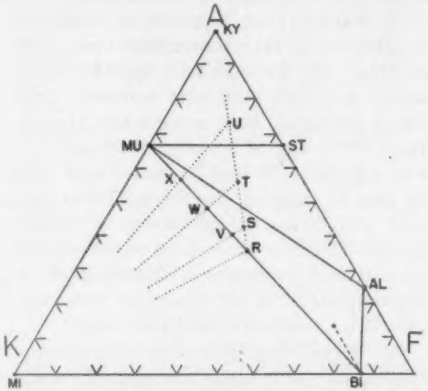

FIGURE 1.—AKF DIAGRAM: STAUROLITE-KYANITE SUBFACIES
KY—kyanite, MU—muscovite, ST—staurolite, AL—almandine, MI—microcline, BI—biotite

sillimanite, almandine, and orthoclase. Its production, under conditions of rising temperature, from the muscovite-biotite schist represented by R of Figure 1 would be a case of truly isochemical metamorphism, and R′ would be a rock of higher metamorphic grade than R.

Regional metamorphism is in many cases accompanied by chemical changes. For example, one of the processes involved in "granitization" may be represented (in a highly simplified way) as an addition of K_2O to the system. On Figure 1, this is shown as a progressive shift of composition toward the K corner of the diagram. Rocks having compositions S, T, U would come to have compositions V, W, X respectively; and almandine, staurolite, and kyanite would be replaced by micas. The line muscovite-biotite represents a critical composition where K_2O is just sufficient to convert all Al_2O_3 into micas; further addition of K_2O would result in the formation of microcline. The appearance of microcline thus marks the final, rather than the initial, stage in the "granitization" of aluminous schists.

These relations may explain some of the irregularities encountered in the mapping of metamorphic zones. The failure or repetition of certain zones may be due to the chemical composition of the sediments. This possibility has long been recognized in the case of staurolite

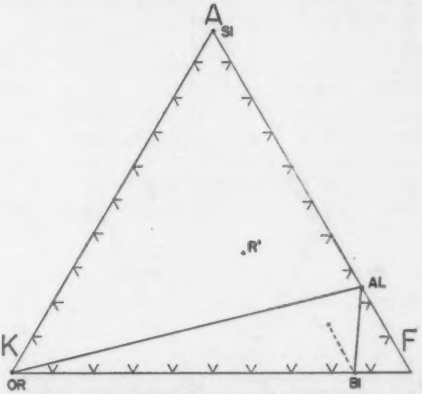

FIGURE 2.—AKF DIAGRAM: SILLIMANITE-ALMANDINE SUBFACIES
SI—sillimanite, AL—almandine, OR—orthoclase, BI—biotite

(Harker, 1939, p. 225); another case has recently been described by Freedman (1950), where the kyanite zone is lacking, staurolite extending to the border of the sillimanite zone.

The sillimanite isograd is generally accepted as marking the transition to a higher grade of metamorphism. Here, however, other considerations arise: the shifting of the phase boundaries from the positions shown in Figure 1 to those shown in Figure 2 involves: (1) the inversion of kyanite to sillimanite; (2) the inversion of microcline to orthoclase; (3) the decomposition of muscovite to form sillimanite and orthoclase; (4) the decomposition of staurolite to form sillimanite and almandine. It is unlikely that all these changes occur simultaneously (Billings, 1950), and there may well be one or more intermediate subfacies between these two. Ch'ih (1950) has described an assemblage containing kyanite and orthoclase (Fig. 3a); Billings (1950) proposes a sillimanite-muscovite subfacies (Fig. 3b); Turner (1948, p. 82) suggests that the assemblage kyanite-almandine-muscovite is stable at somewhat higher temperatures than the assemblages containing staurolite which it replaces (Fig. 3c).

The last two subfacies are difficult to recognize except in Al_2O_3-rich rocks; both show a muscovite-biotite-almandine triangle within which rocks would be indistinguishable from those of similar composition in the staurolite-almandine sub-facies (Fig. 1). Figures 2 and 3a show the conversion of muscovite to orthoclase and sillimanite (kyanite). This is probably a more significant index for the highest grade of metamorphism than the change of kyanite or staurolite to sillimanite, which may occur at lower temperatures, or perhaps only under certain physical or chemical conditions.

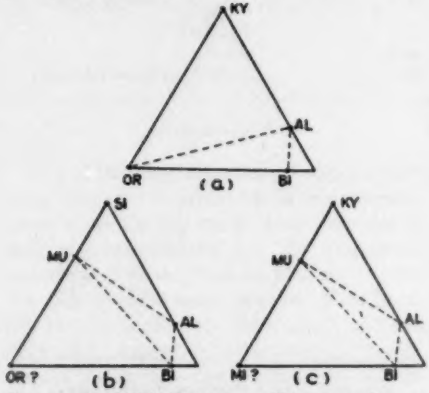

FIGURE 3.—AKF DIAGRAMS: POSSIBLE INTERMEDIATE SUBFACIES
(a) "kyanite-orthoclase subfacies", (b) "sillimanite-almandine subfacies", (c) "kyanite-almandine subfacies"

PHILADELPHIA REGION AS A TEST CASE

These theoretical considerations may profitably be applied to the study of the pelitic rocks of the Wissahickon formation in the Philadelphia region. In the part of the region dealt with here (Figs. 17–22), the mica schists and mica gneisses belong to the amphibolite facies and exhibit the full range of "index" minerals: chlorite, biotite, almandine, staurolite, kyanite, and sillimanite. The area appears suitable for detailed study of the relation of these minerals to the chemical character of the rocks and to the physical conditions of regional metamorphism. Materials for such a study have been accumulated, during more than 60 years, in the collections of Bryn Mawr College and in published reports on the geology of eastern Pennsylvania.

My own work was begun with the idea of metamorphic zones in mind. A preliminary map prepared in 1936, based on comparatively few, widely scattered data, suggested an increase in the grade of regional metamorphism toward the southeast, with the sillimanite isograd roughly parallel to a belt containing numerous large igneous intrusives. Maps recently published by Weiss (1949) and McKinstry (1949) give the same impression. In 1945 I resumed work with the idea of mapping the isograds more precisely, and, finding the distribution of "index" minerals difficult to account for on the hypothesis that each represents a different grade of metamorphism, was led to consider more carefully the implications of the facies concept.

The Wissahickon formation shows great chemical diversity: within the limits of a single quarry or other large exposure, highly micaceous schists (locally rich in almandine, staurolite, or kyanite) may be seen alternating with and grading into massive gneisses composed chiefly of quartz and feldspar with little mica. Even if "zonal" mapping is based on the micaceous schists alone (as representing the most aluminous sediments), petrographic study shows that these exhibit a wide range in the relative proportions of muscovite and biotite, or of quartz and feldspars, indicating that they are far from uniform in chemical composition.

An additional factor making for chemical diversity is the "granitization" which has affected some parts of the region. This was first investigated by Postel (1940) and has recently been discussed in more detail by Ch'ih (1950). These granitized rocks, formerly believed to be igneous intrusives, have been mapped as "granite gneiss" (Bascom, 1909a) or "granodiorite" (Weiss, 1949).

Interpretation of the metamorphic history therefore involves discussion of several intricately related questions:

1. What is the relative importance of chemical composition and of physical (temperature-pressure-stress) conditions in the formation of the aluminosilicate "index" minterals?

2. Is the hydrothermal activity ("granitization") which has affected many of the rocks an

TABLE 1.—CHEMICAL ANALYSES, A–J

	A	B	C	D	E	F	G	H	J
SiO_2	73.68	56.40	59.39	60.33	74.24	75.04	75.15	73.59	79.60
Al_2O_3	12.49	19.76	16.38	20.85	13.71	12.59	13.15	11.37	9.48
Fe_2O_3	2.10	4.35	4.82	3.59	2.01	2.45	2.70	2.82	1.77
FeO	2.22	4.40	1.96	4.47	—	—	—	—	1.49
MgO	2.04	3.11	2.11	2.07	1.09	0.38	0.68	0.77	0.76
CaO	0.56	0.09	7.33	1.82	1.68	1.44	1.31	1.62	0.72
Na_2O	2.97	5.82	0.53	1.38	0.60	1.62	1.95	2.07	1.83
K_2O	2.91	1.27	1.64	2.84	4.84	3.98	5.21	4.65	1.54
H_2O	1.34	3.37	1.92	2.78	0.48	0.50	0.38	0.74	1.66
TiO_2	0.81	1.05	1.24	1.41	0.36	0.50	0.28	1.80	0.71
P_2O_5	0.12	0.37	0.13	0.28	0.26	0.31	0.22	0.07	0.19
MnO	0.18	tr	—	—	—	—	—	tr	0.67
Fe_7S_8	—	—	3.06	—	—	—	—	—	—
	101.42	99.99	100.51	101.82	99.27	98.81	101.03	99.50	100.42

Analyses A–J by F. A. Genth, Jr. (Hall, Geology of Philadelphia County, vol. C6, 1881; vol. 00, Catalogue of the Geological Museum, Part II, 1880, gives additional information as to localities). Since no thin sections of these specimens are available, Genth's descriptions are given *verbatim*, with remarks based on my own study of thin sections which are probably from the same localities.

A. (C6, p. 108, No. 4708) "*Gneiss* . . . Fine-grained mixture of quartz and feldspar, with a small quantity of little scales of muscovite and biotite."
Republished by Bascom (1905, p. 304, No. 6; 1909a, p. 4, No. 3; 1909b, p. 4, No. 3); by Weiss (1949, p. 1698, col. 7).
None of the later publications give the locality clearly and the original report (Hall, 1880, p. 16–19) has been misinterpreted. From the context it is clear that the whole series of traverses along Neshaminy Creek was made from north to south (i.e., *down* stream); but the list gives all measurements as "above" the starting point. Genth (Hall, 1881, p. 108–110) evidently understood this; but Bascom consistently refers to this locality as "below" the next (listed as B, below), which wrongly suggests that it is farther down stream (south). The traverse may easily be traced on the Burlington (Pa.–N.J.) quadrangle, from (what is now) Parkland to Hulmeville; station 4708 (analysis A) is about 0.9 mi. north of the bridge at Hulmeville, and station 4771 (analysis B) about 0.2 mi. north of the bridge. Weiss (1949, p. 1698; 1712; Pl. 1) is therefore mistaken in identifying analysis A with locality 45–T26, which is about a mile southwest of Neshaminy Falls on the Burlington quadrangle.

B. (C6, p. 109, No. 4771) "*Gneiss* . . . Irregular bands of quartz and feldspar, inclosed between scales and sheet-like aggregations of scales of muscovite and biotite."
Republished by Bascom (1905, p. 304, No. 4; 1909a, p. 4, No. 1; 1909b, p. 4, No. 1).
For note on locality, see preceding analysis A.
Specimen 29–54 has almandine and a trace of kyanite.

C. (C6, p. 121, No. 5772) "*Gneiss* . . . White feldspar, quartz and biotite, the latter in brown and brownish black scales in great quantity, also small quantities of pyrrhotite and magnetite."
East side of Pennypack Creek at Vereeville bridge.
Specimens 52–120, 121 contain almandine and microcline.

D. (C6, p. 122, No. 5929) "*Hydromica schist (?)* . . . A grey micaceous rock. The particles are intimately interwoven and contain grains of garnet, white feldspar and very little magnetite."
Republished by Weiss (1949, p. 1698, col. 5) giving "Florence Bascom's notes" as source.
North of Jenkintown Junction on west side of Tacony Creek.
Specimen 47-P 268 contains a little staurolite.

E. (C6, p. 122–123, No. 6101) "*Gneiss* . . . Finely granular greyish-white rock, made up of feldspar and quartz, small scales of biotite and a little magnetite."
Republished by Weiss (1949, p. 1711, col. 4) giving "Florence Bascom's notes" as source and erroneously listing this as hornblende gneiss; it is evidently a mica gneiss, similar to the three following; all four are from Frankford.
Specimen 52–218 contains microcline.

F. (C6, p. 123, unnumbered) "*Gneiss* . . . It is of fine granular texture, composed of greyish-white quartz and white feldspar; disseminated through the mass are spots of fine scaly biotite, frequently inclosing particles of red garnet."
Frankford. Specimen 52–218 contains microcline.

G. (C6, p. 123–124, unnumbered) "*Gneiss* . . . "It is of fine-grained texture and a mixture of feldspar and quartz, of a yellowish-greyish-white color, and disseminated through it are dark stripes, produced by fine scales of biotite."
Frankford. Specimen 52–218 contains microcline.

TABLE 1.—(*Continued*)

H. (C6, p. 124, unnumbered) "*Granite* . . . Fine-grained, almost wholly composed of quartz and feldspar with a little mica."
Frankford. Specimen 52–218 contains microcline.

J. (C6, p. 135, unnumbered) "*Gneiss* . . . A very fine-grained mixture of quartz and feldspar with a little fine scaly mica; easily breaking into flat angular pieces."
Republished by Knopf and Jonas (1929, p. 31, No. 3); by Cloos and Hietanen (1941, p. 142, No. III).
Narberth Station.
Specimen 36-Nar 2, which closely resembles Genth's description, contains orthoclase and kyanite. (This analysis is shown on both Fig. 4 and Fig. 5.)

extraneous factor, to be isolated if possible and disregarded, or is it a controlling factor in the regional metamorphism?

3. If rocks of different metamorphic grade occur in the region, is the relation among them one of progressive or of retrogressive metamorphism?

This paper is a record of the investigation of these questions; I hope later to deal in a similar way with the hornblende schists and gneisses of the region.

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CHEMICAL CHARACTER OF THE ROCKS

The mica schists and granitized schists ("granites", "granite gneisses", or "granodiorites") are represented by 22 chemical analyses (Tables 1, 2). All except 3 have been published before; but some appear to have been forgotten and others have been incorrectly cited. Localities are shown by corresponding letters or numbers on Figure 22.

These 22 analyses have been plotted on AKF diagrams (Figs. 4, 5). Analyses 1–13 have been corrected as Eskola (1939, p. 347) recommends, by micrometric analysis for magnetite and sphene, as well as for apatite. Analyses A–J, for which no thin sections are available, have been corrected only for apatite, to the amount indicated by the P_2O_5 in the analyses.

The theoretical assemblages may now be compared with the actual mineral compositions of the rocks. In Table 1, no quantitative modes are given, since no thin sections are available. I have not recalculated modes for these analyses because: (1) The AKF diagrams show clearly the ratios between Al_2O_3 and ($CaO + Na_2O + K_2O$) which determine whether the minerals almandine, staurolite, kyanite, or sillimanite are likely to be formed; (2) In the absence of data on the amount of these minerals, as well as the amount of magnetite or ilmenite and the composition of the biotite, recalculated modes would give a specious impression of accuracy. The modes given in Table 2, were obtained by micrometric analyses.

Table 3 summarizes a comparison of these modes with the theoretical assemblages shown on Figures 4 and 5. The correspondence is in general good, and some of the apparent discrepancies can probably be explained:

Analysis C. If both almandine and microcline are present, the rock (52–120, 121) is perhaps in disequilibrium; if reactions had been completed, it might well belong in 3a.

Analysis 1. This has been omitted from the AKF diagram, since it shows a small deficiency of Al_2O_3; specimen 44–48a is a highly quartzitic schist.

Analyses 4 and 5. On the AKF diagram these lie so close to the mica field that perhaps little kyanite and staurolite or microcline would be expected.

On the whole, the AKF diagram (Fig. 4) appears to give a satisfactory representation of the mineral assemblages of mica schists, and strongly suggests that the production of microcline, or of kyanite, staurolite, and almandine is largely controlled by chemical composition in rocks of the staurolite-kyanite subfacies.

In the sillimanite-almandine subfacies (Fig. 5), relations are less easy to interpret. It is noticeable that analysis 10 falls in field II and contains only a trace of kyanite; while the

others, which are in field I, all contain more kyanite or sillimanite.

Only one of these specimens, however, appears to be in real equilibrium—analysis 13 (orthoclase-biotite-sillimanite). Analysis 12 (muscovite-biotite-sillimanite-almandine) may represent a rock of an intermediate sillimanite-muscovite subfacies (Fig. 3b), though it is not in the sillimanite field. Analysis 11 (orthoclase-muscovite-biotite-kyanite), analyses 10 and probably J (orthoclase-biotite-kyanite) may belong to an intermediate orthoclase-kyanite subfacies (Fig. 3a), though again the presence of muscovite in analysis 11 would indicate disequilibrium.

Since these last 4 analyses (J and 10–12) all lie within the muscovite-biotite-almandine triangle, we may conclude that assemblages with sillimanite, orthoclase, or orthoclase-kyanite are formed under physical conditions different from those of the staurolite-kyanite subfacies, and probably represent a higher grade of metamorphism.

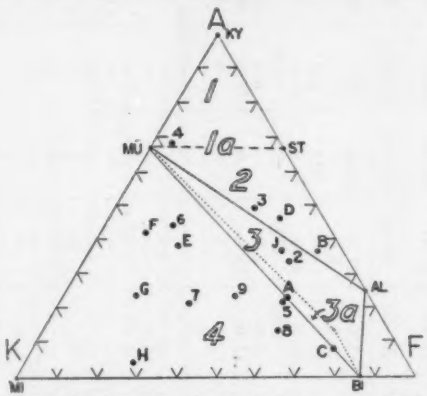

FIGURE 4.—CHEMICAL ANALYSES PLOTTED ON AKF DIAGRAM FOR STAUROLITE-KYANITE SUBFACIES

(See Fig. 1.) Large outline numbers indicate stable mineral assemblages; smaller letters and numbers refer to chemical analyses listed in Tables 1 and 2. 1a. Kyanite-almandine-muscovite (biotite, plagioclase, quartz). This assemblage, according to Turner (1948, p. 82) replaces assemblages with staurolite (1, 2) at slightly higher temperatures.

1. Kyanite-staurolite-muscovite (biotite, plagioclase, quartz).
2. Staurolite-almandine-muscovite (biotite, plagioclase, quartz).
3. Almandine-muscovite-biotite (plagioclase, quartz).
3a. Muscovite-biotite (plagioclase, quartz).
4. Muscovite-biotite-microcline (plagioclase, quartz).

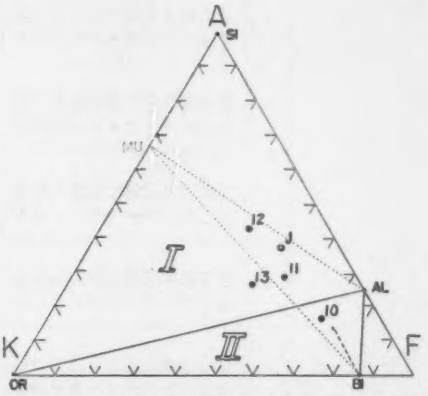

FIGURE 5.—CHEMICAL ANALYSES PLOTTED ON AKF DIAGRAM FOR SILLIMANITE-ALMANDINE SUBFACIES

(See Figs. 2, 3.) Large outline numbers indicate stable mineral assemblages; smaller letters and numbers refer to chemical analyses listed in Tables 1 and 2.

I. Sillimanite-almandine-orthoclase (biotite, plagioclase, quartz).

II. Almandine-biotite-orthoclase (plagioclase, quartz).

Dotted lines delimit the field muscovite-biotite-almandine, which may be a stable assemblage in an intermediate sillimanite-muscovite or kyanite-almandine subfacies (see Fig. 3b, c).

SIGNIFICANT MINERAL ASSOCIATIONS

Classification of Specimens

The thin sections used for petrographic study have been classified as shown in Table 4. The assemblages of the staurolite-almandine subfacies (Table 4) have been numbered as in Figure 4. Assemblage 1a has been included here, although it may represent a transitional subfacies (Fig. 3c). Assemblages 1 and 2 have been grouped together simply because so few specimens are available. Rocks containing almandine and micas, or micas only (3 and 3a) have also been assigned to this subfacies, although some of them may belong to a subfacies transitional to the sillimanite-almandine subfacies (Fig. 3, b or c). The only rocks which appear to be in disequilibrium are those which contain microcline as well as almandine; these

TABLE 2.—CHEMICAL ANALYSES, 1–13

Chemical analyses	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	81.90	69.85	67.40	78.48	69.94	70.48	72.07	66.65	69.70	68.01	72.38	60.56	75.99
Al_2O_3	8.59	13.74	16.61	11.56	14.12	17.84	15.13	15.66	15.33	13.86	13.25	18.18	12.37
Fe_2O_3	0.47	1.82	3.76	2.57	1.28	0.99	0.71	1.01	0.91	0.63	0.69	5.30	0.78
FeO	2.10	3.08	3.58	0.69	4.10	0.68	1.69	3.44	2.29	5.28	3.50	3.73	1.91
MgO	0.13	2.34	1.36	0.12	1.61	0.28	0.55	1.31	0.90	2.22	1.93	2.18	1.03
CaO	1.67	0.78	0.58	0.40	1.29	1.89	1.89	3.05	2.48	1.03	0.75	0.40	1.24
Na_2O	2.98	2.34	0.87	1.41	2.31	3.58	2.59	3.43	3.28	3.77	1.92	1.24	2.87
K_2O	1.16	2.57	3.25	2.26	3.66	3.89	4.93	3.61	3.58	2.65	3.06	4.50	2.74
H_2O	0.62	1.36	1.84	1.97	1.46	0.52	0.25	0.70	0.72	0.99	1.48	2.23	0.42
TiO_2	0.60	1.46	0.75	0.61	0.75	0.28	0.26	0.78	0.61	1.19	0.69	1.09	0.54
P_2O_5	—	0.11	—	—	—	—	—	0.21	0.10	0.19	0.13	0.17	0.17
MnO	0.03	0.08	0.11	0.03	0.08	0.03	0.04	0.07	0.04	0.06	0.08	0.20	0.04
	100.25	99.53	100.11	100.10	100.60	100.46	100.11	99.92	99.94	99.88	99.86	99.78	100.10
Mode (vol. %)	44–48a	44–9	44–42	52–100	52–15	52–46	28–230	28–231	23–22	23–125	36–576 S	23–107	47–P 449
Quartz	72.0	44.0	48.0	61.0	48.0	32.0	31.0	45.0	27.0	28.0	56.0	40.0	46.0
Microcline	—	—	—	—	—	17.0	17.0	8.0	7.0	—	—	—	—
Orthoclase	—	—	—	—	—	—	—	—	—	0.5	4.0	—	25.0
Plagioclase	21.0	20.0	3.0	4.0	27.0	38.0	42.0	31.0	48.0	45.0	11.0	2.0	17.0
(% an)	(16)	(10)	(25)	(23)	(22)	(24)	(23)	(24)	(30)	(28)	(22)	(8)	(24)
Muscovite	—	19.0	27.0	21.0	7.0	8.0	—	3.0	1.0	—	6.0	29.0	—
Biotite	6.2	14.0	8.4	12.0	17.0	2.0	8.0	11.0	15.0	26.0	21.0	19.0	6.5
Almandine	0.1	tr	12.0	—	0.7	—	—	—	—	—	tr	2.0	—
Staurolite	—	—	tr	—	—	—	—	—	—	—	—	—	—
Kyanite	—	—	—	—	—	—	—	—	—	tr	2.0	—	—
Sillimanite	—	—	—	—	—	—	—	—	—	—	—	5.0	5.0
Epidote	0.1	—	—	—	—	2.5	1.0	0.6	—	—	—	—	—
Sphene	—	—	—	—	—	0.5	0.5	1.2	1.4	—	—	—	—
Tourmaline	—	—	—	—	—	—	—	—	—	—	—	0.6	—
Chlorite	—	1.5	—	—	—	—	—	—	—	—	—	tr	—
Apatite	0.1	tr	tr	—	—	tr	—	0.2	tr	0.1	—	tr	—
Magnetite	0.5	1.5	1.4	1.0	0.3	—	0.5	—	0.6	0.4	tr	2.4	0.5
Hematite	—	—	—	0.8	—	—	—	—	—	—	—	—	—
Zircon	—	—	0.2	0.2	—	—	—	—	—	—	—	—	—

Analyses 1–13 were made in connection with work done at Bryn Mawr College:

Hematite														
Zircon	—	—	0.2	0.2	—	—	—	—	—	—	—	—	—	—

Analyses 1–13 were made in connection with work done at Bryn Mawr College:

1. Cloos and Hietanen (1941, p. 142, No. 1).
   Analyst, E. M. Hardy, Bryn Mawr College.
2. Cloos and Hietanen (1941, p. 142, No. I).
   Analyst, O. C. Bates, Bryn Mawr College.
3. Cloos and Hietanen (1941, p. 142, No. II).
   Analyst, R. Stoddard, Bryn Mawr College.
4. Weiss (1949, p. 1698, col. 1).
   Analyst, P. Vipond, Bryn Mawr College.
5. Weiss (1949, p. 1698, col. 6).
   Analyst, M. Quin, Bryn Mawr College.
6. Weiss (1949, p. 1716, col. 1).
   Analyst, P. Vipond, Bryn Mawr College.
7. Ch'ih (1950, p. 928, No. II).
   Analyst, P. Vipond, Bryn Mawr College.
8. Ch'ih (1950, p. 928, No. IV).
   Analyst, E. H. Oslund, University of Minnesota.
9. Ch'ih (1950, p. 928, No. III).
   Analyst, E. H. Oslund, University of Minnesota.
10. Ch'ih (1950, p. 928, No. V).
    Analyst, E. H. Oslund, University of Minnesota.

11, 12, 13. Analyst, S. S. Goldich, University of Minnesota.

TABLE 3.—COMPARISON BETWEEN ACTUAL AND THEORETICAL MINERAL ASSEMBLAGES

Classification	A	B	C	D	E	F	G	H	J
From Table 1	3a	1a	3 ? 4 ?	2	4	3 ? 4 ?	4	4	3a ? I ?
From AKF diagrams (Figs. 4 and 5)	3a	1a	3a	2	4	4	4	4	3 ? I ?

	1	2	3	4	5	6	7	8	9	10	11	12	13
From Table 2	3	3	2	3a	3	4	4	4	4	II ?	I ?	I ?	I
From AKF diagrams (Figs. 4 and 5)	—	3	2	1 ? 1a ?	4	4	4	4	4	II	I	I	I

TABLE 4.—GROUPING OF SPECIMENS FOR PETROGRAPHIC STUDY

STAUROLITE-KYANITE SUBFACIES

Assemblage	1a	1-2	3	3a	4
Number of specimens	34	16	115	96	129
Kyanite	X	e			
Staurolite		X			
Almandine	e	e	X		
Micas	e	e	e	X	e
Microcline			d		X
Plagioclase, Quartz	e	e	e	e	e

SILLIMANITE-ALMANDINE SUBFACIES

Assemblage	A (23 specimens)						B (29 specimens)					C (62 specimens)						
Number of specimens	3	4	7	1	6	2	9	6	1	9	4	28	2	13	9	2	6	2
Sillimanite	X	X	X	X			X	X	X			X	X	X	X	X		
Kyanite			d	d	u			d	d	u	u			d	d			
Staurolite															u	u		
Almandine	e	e	e	e	e		e	e	e	e	e	e	e	e	e	e		
Biotite	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e
Muscovite							u	u	u	u	u	u	u	u	u	u	u	u
Orthoclase		X	X	X	X	X	X	X	X	X	X						X	X
Microcline				d					d		d		d					d
Plagioclase, Quartz	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e

X—index mineral for the assemblage.
e—mineral commonly present in equilibrium.
d—mineral present in disequilibrium.
u—mineral present: equilibrium relations uncertain.

have been assigned to assemblage 3, although if equilibrium had been attained they would perhaps belong in 3a or 4.

The areal distribution of these specimens is shown on Figure 17.

Rocks containing orthoclase or sillimanite are more difficult to classify. The data are given in some detail (Table 4). Evidently some of these are disequilibrium assemblages, notably microcline-orthoclase (8 specimens), kyanite-

sillimanite (28 specimens), staurolite-(kyanite-) sillimanite (11 specimens). Whether the association kyanite-orthoclase (19 specimens) is stable is questionable. If it is, we might be justified in defining an intermediate kyanite-orthoclase subfacies (Fig. 3a); but 13 of these specimens contain muscovite, which seems to indicate that equilibrium has not been attained, and the remaining 6 specimens are too few to form the basis for any decision.

In Table 4, the inversions kyanite $\rightleftharpoons$ sillimanite and microcline $\rightleftharpoons$ orthoclase have been disregarded, in order to bring out the possible phase relations between these minerals and muscovite.

Group A (23 specimens), without muscovite, can presumably be assigned to the sillimanite-almandine subfacies as defined by Turner (1948, p. 85) or to an intermediate orthoclase-kyanite subfacies (Fig. 3a).

All the remaining specimens contain muscovite and may be interpreted in two ways:

(1) Some of them, at least, may be equilibrium rocks representing transitional subfacies. The largest group (28 specimens) shows the association sillimanite-muscovite, which supports Billings' (1950) contention that this subfacies exists (Fig. 3b). With these are placed 8 specimens with orthoclase-muscovite and, tentatively, specimens with kyanite or staurolite in addition to sillimanite-muscovite (Group C, 62 specimens).

(2) The presence of muscovite may, however, indicate disequilibrium, with sillimanite or orthoclase or both in process of alteration by hydrothermal solutions. This is almost certainly the case with rocks which contain muscovite together with both orthoclase and sillimanite (kyanite), which I have placed in Group B (29 specimens). Such associations might be stable if formed under conditions where water was limited in amount, muscovite forming only to the amount of the available H_2O; but I believe this hypothesis may be rejected, since other evidence indicates that water was present, probably in excess, during the metamorphism.

The final grouping of specimens in Table 4 thus represents the following conclusions:

(a) There is evidence for the existence in this region of a sillimanite-almandine subfacies (in the narrowest sense—Group A); there is dubious evidence for a kyanite-orthoclase subfacies (Group A), and rather better evidence for a sillimanite-muscovite subfacies (Group C). Group B represents a transition between Groups A and C.

(b) Whatever the true number of subfacies, there is much evidence of disequilibrium. I shall therefore, for the present, retain the term sillimanite-almandine subfacies for all three groups, considered as a whole, and reconsider the question of intermediate subfacies in a later section.

The areal distribution of these specimens is shown on Figure 18.

In Table 4 the accessories (magnetite, ilmenite, apatite, zircon) and the minerals epidote, sphene, chlorite, and tourmaline have been disregarded, although (as will be shown) the distribution of the last four minerals offers important evidence as to the metamorphic history of the region.

Descriptions of Minerals

Brief notes on some individual minerals follow, and their distribution is shown (Fig. 7) in terms of percentage of rocks in each group in which each mineral appears. This method of presentation facilitates rapid comparison of the different groups, and the distribution patterns bring out tendencies, perhaps not very striking in any one group, which suggest important relationships. Some of these tendencies might otherwise escape notice, since there are many variable features and the processes which they record have in many cases failed to reach completion.

Orthoclase (Fig. 7a) has been carefully studied by Ch'ih (1950, p. 927); it is commonly perthitic and may grade into or be bordered by microcline.

Microcline (Fig. 7a) shows typical grid twinning and no unusual optical properties.

Myrmekite (Fig. 7a) (quartz-plagioclase intergrowth) is common, associated with either orthoclase or microcline; in a very few rocks, it occurs without potash feldspar.

Plagioclase (Fig. 8a) ranges in composition from albite to calcic andesine, but is most commonly oligoclase; it is generally twinned, but in some rocks untwinned oligoclase is difficult to distinguish from quartz. Zoning (Fig. 7b) is of the "normal" type (not uncommon) or of the "reverse" type (rare); compositional differences

between rim and core rarely exceed 2–3 per cent.

Biotite (Fig. 7c, d) occurs in all the rocks of the sillimanite-almandine subfacies and in 97 per cent of those of the staurolite-kyanite subfacies. It exhibits a considerable range of color and pleochroism, undoubtedly indicating a considerable range in chemical composition. Criteria based on color alone are necessarily qualitative and somewhat subjective, but I have adopted 4 categories:

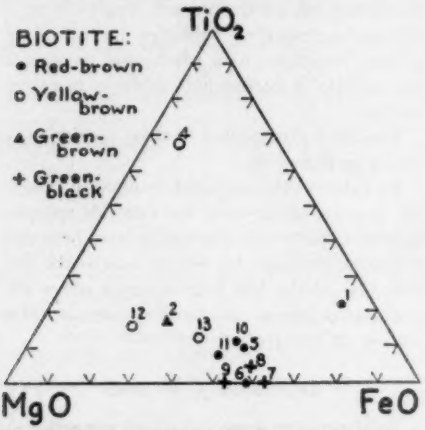

FIGURE 6.—TRIANGULAR DIAGRAM SHOWING RELATION BETWEEN COLOR OF BIOTITE AND CONTENT OF TiO_2, MgO, AND FeO
Numbers refer to chemical analyses listed in Table 2.

Red-brown:	*X*, pink-buff to orange-yellow *Y* = *Z*, red-brown to orange-brown
Yellow-brown:	*X*, pale buff to yellow *Y* = *Z*, dull yellow-brown
Green-brown:	*X*, pale yellow to yellow-green *Y* = *Z*, green-brown or olive brown
Green-black:	*X*, yellow-green or yellow-brown *Y* = *Z*, dark green to black (opaque on basal sections)

No chemical analyses of these biotites are available, but some indication of their probable composition may be gained by the rather unsatisfactory method of recalculation from rock analyses. Analyses 1–13 have been corrected by micrometric analyses for the amount of TiO_2 in sphene and of FeO in magnetite, and analysis 12 also for the amount of FeO in almandine. (Analysis 3 has been omitted, since this last correction results in a large deficit of FeO.) Assuming that the remaining TiO_2, FeO, and MgO are combined in biotite, these oxides (as molecular per cent recalculated to 100) are plotted on Figure 6, symbols indicating the color of the biotite in each specimen. The results show a fair agreement with the findings of Hall (1941) that reddish or orange biotites are richer in TiO_2 than greenish biotites. It also appears that deeper color (red-brown or green-black) is due to preponderance of FeO over MgO. The agreement is better if only analyses 7–13 be considered; these are probably more reliable than the rest, since material for chemical analyses and thin sections used for micrometric analyses were taken from the same specimens.

Muscovite (Fig. 7d). White mica is present in nearly 80 per cent of the rocks of the sillimanite-almandine subfacies and nearly 90 per cent of those of the staurolite-kyanite subfacies. Some white mica shows an unusually small optic angle and has been described as phengite (Berman, 1938; Postel, 1940). One chemical analysis (Postel and Adelhelm, 1944) indicates that this mica is neither a true phengite nor a normal muscovite; it is rather high in Fe_2O_3 and Na_2O. The material discussed by all these authors is from "granitized" or "hydrothermally formed" rocks and the unusual properties may be due to a replacement origin. Optical properties alone do not permit separation of this from "normal" muscovite in thin section, since micas are commonly bent or in overlapping scales and give unreliable values for *2V*. I have therefore called all white mica muscovite.

In some rocks, white mica (probably in most cases the "phengitic" type referred to above) forms a symplectitic intergrowth with plagioclase (Fig. 7e).

Epidote (Fig. 7e) is most abundant in rocks containing myrmekite and muscovite symplectite, and it too may form intergrowths with plagioclase. Epidote is pale green with faint pleochroism and high birefringence; not rarely zoned or with cores of (altered) allanite. Weiss (1949, p. 1695) described as zoisite "minute,

rounded, pink inclusions in quartz and plagioclase". These objects, which are almost completely restricted to quartz, are quite common in the coarser-grained schists in the northwestern part of the region. They are probably minute cavities, perhaps containing liquid: their index of refraction is much lower than N_0 of quartz. Zoisite (or clinozoisite) forming as "saussurite" at the expense of plagioclase is extremely rare.

Sphene (Fig. 7e), usually found in association with epidote or green-black biotite, is brown, with typical high birefringence and dispersion, occuring in subhedral grains or granular aggregates.

Chlorite (Fig. 7f) is commonly associated with or interleaved with biotite, but is nowhere abundant enough to be considered a major constituent. It is weakly pleochroic, with X nearly colorless to pale green, $Y = Z$ pale green to blue-green; birefringence low, with blue tints characteristic of penninite.

Tourmaline (Fig. 7f) occurs as small euhedral prisms or as larger irregular poikiloblastic grains; it is strongly pleochroic, with O dark gray-blue to blue-green, E colorless to pale pink or violet. Zoned crystals have blue cores and green or brownish-blue rims. A much rarer type has O dark yellow-brown and E pale buff or yellow.

Mineral Associations as Indices of Metamorphic Processes

Indices of Granitization

A general similarity in the distribution patterns of microcline, myrmekite (Fig. 7a), normally zoned plagioclase (Fig. 7b), muscovite symplectite, epidote, sphene (Fig. 7e), and green-black biotite (Fig. 7d) suggests a common origin. Indeed the typical rock in which several or all of these minerals occur together is a microcline gneiss or "granodiorite", already referred to as a product of "granitization".

May we then accept these as "index" minerals diagnostic of "granitization"? This question involves discussion of several points:

1. Postel (1940), who first pointed out the importance of hydrothermal changes in the formation of these rocks, nevertheless believed some of the "granodiorites" in the area he studied to be true magmatic rocks; this opinion was based chiefly on cross-cutting relations observed in the field. Ch'ih (1950) studied the granitization process in more detail and made fabric analyses of typical rocks; she concluded that even the apparently intrusive "dikes" have been formed by replacement of mica schists. My own opinion coincides with Ch'ih's; but in any case, "granitic magma", if present, was certainly accompanied by mobile watery solutions, and magmatic rocks would have been "stewed in their own juice" until they finally reached equilibrium under the same physical conditions which were controlling the recrystallization of the adjacent granitized schists.

2. Another possibility is that some of the schists were originally arkosic sands or volcanic tuffs which contained enough K_2O to form microcline. Actually microcline is accompanied by one or more of the other minerals here taken as evidence of hydrothermal processes in all but 7 of the specimens found. These seven occur in close proximity to others containing myrmekite, symplectites, etc., and therefore seem unlikely to have had a different origin.

3. In discussions of "granitization", emphasis tends to be placed on the introduction of K_2O, since it is the production of microcline which transforms a mica schist into a "granite-like rock"; but, as already noted (Fig. 1), the appearance of microcline marks the final rather than the incipient stages of "granitization". The early stages, characterized by the formation of micas at the expense of aluminosilicates, are more difficult to identify except where replacement textures, symplectites, etc., can be recognized. Moreover, as Ch'ih (1950) has shown, Na_2O and probably CaO have been introduced, resulting in such features as myrmekite, zoning of plagioclase, and production of epidote and sphene; and FeO, MgO and TiO_2 have been redistributed, resulting in changes in the character of the biotite. These processes can be identified in some rocks in which the concentration of K_2O is not great enough to produce microcline.

I therefore adopt as a working hypothesis the assumption that the presence of any of the minerals listed above indicates hydrothermal

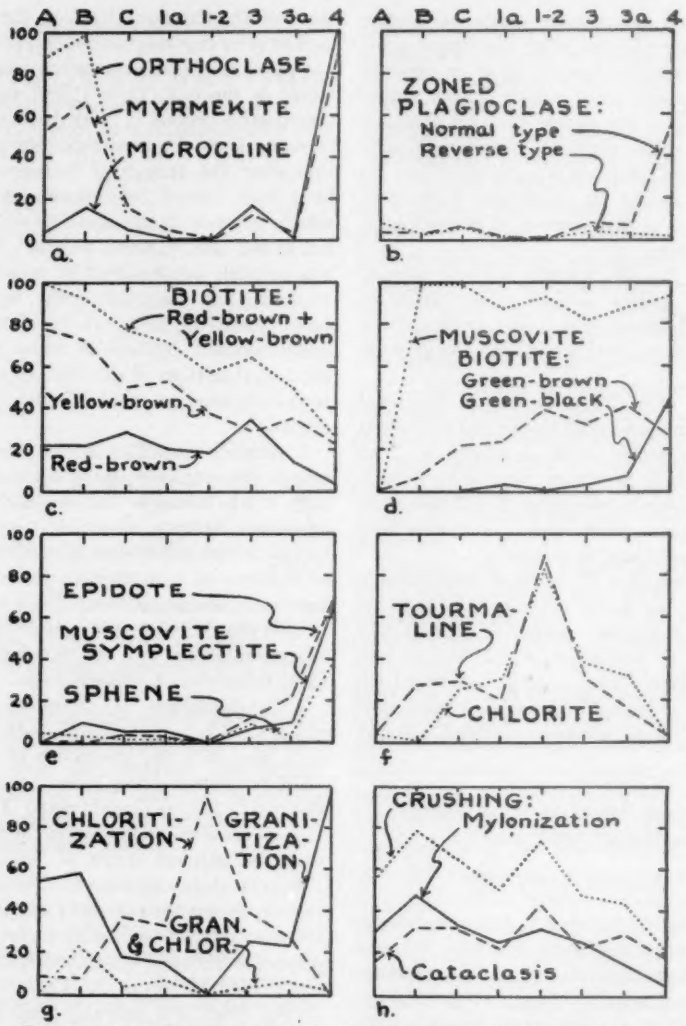

FIGURE 7.—DISTRIBUTION OF SIGNIFICANT MINERALS, OF HYDROTHERMAL CHANGES, AND OF CRUSHING
For groups, see Table 4.

activity. For brevity, I shall refer to rocks containing any of the these minerals as granitized rocks, and to the process as granitization.

4. Ch'ih (1950) suggests that orthoclase is a product of granitization. On the other hand, the association orthoclase-sillimanite is supposed to be stable in rocks formed by high grade regional or contact metamorphism. I have therefore not taken the presence of orthoclase *alone* as an index of granitization; readers who prefer Ch'ih's hypothesis will note that groups A and B (Fig. 7a, g) would then be considered 87–100% granitized.

5. Finally there is the question of the role of muscovite, which might be expected to form during granitization. The distribution pattern of muscovite (Fig. 7d) is not very similar to those already discussed. Evidence is presented

later to show that muscovite is not a suitable index of granitization, since it forms also in the range where chlorite is stable.

Indices of Chloritization

The "sympathetic" distribution of chlorite and tourmaline is obvious (Fig. 7f), and the pattern of these two minerals is for the most part "antipathetic" to those of the minerals regarded as indices of granitization. If tourmaline and chlorite are formed by a continuation of the hydrothermal activity responsible for granitization, they appear to belong to a separate stage of the process. For brevity, I shall refer to this stage as chloritization and to rocks containing chlorite, tourmaline, or both, as chloritized rocks.

Relation Between Granitization and Chloritization

Figure 7g, shows the distribution of rocks as classified into 3 categories:

1. Granitized rocks: containing microcline, myrmekite, plagioclase with normal zoning, muscovite symplectite, epidote, sphene, or green-black biotite.

2. Chloritized rocks: containing chlorite or tourmaline.

3. Rocks containing any of the minerals listed in 1 together with chlorite or tourmaline —i.e., rocks showing evidence of both chloritization and granitization. This class is small, indicating rather little overlapping of the two stages; in the following sections, these rocks are grouped with chloritized rocks.

The areal distribution (Figs. 19, 20) is also significant: chloritization is predominant in the northwestern half of the region, where granitization is subordinate, and is less common in the zone of most intense granitization.

Indices of Regional Metamorphism

Having now eliminated certain minerals and mineral associations as indicative of hydrothermal activity, can we find among the remaining minerals any which show progressive changes indicative of increasing (or decreasing) "pure" regional metamorphism—i.e., minerals the occurrence or character of which appears to be controlled largely by temperature gradient, by the intensity of regional stress, or by some combination of the two?

The aluminosilicates almandine, staurolite, kyanite, and sillimanite, generally considered index minerals in this sense, will be discussed later. We may first examine, from this point of view, the data already presented for plagioclase, biotite and muscovite.

Plagioclase (Fig. 8a) is represented in terms of the average composition (per cent anorthite) for each group; this is slightly higher for rocks of the sillimanite-almandine subfacies than for rocks of the staurolite-kyanite subfacies. The difference is small, but is consistent with the generally held idea that more calcic plagioclases are characteristic of higher grades of metamorphism. In the staurolite-kyanite subfacies, however, the graph is so irregular that further study is needed.

Biotite (Fig. 7c, d) shows several points of interest:

1. The graphs for red-brown and yellow-brown biotites (Fig. 7c), are markedly antipathetic, suggesting that they may be simply alternative varieties of the same type of biotite, both containing more TiO_2 than the greenish biotites, but differing in the relative proportions of MgO and FeO (*see* Fig. 6).

2. If red-brown and yellow-brown varieties be taken together, a marked trend is observable, decreasing from left to right. This distribution is in keeping with the observations of Tilley (1925) and of Phillips (1930) that reddish biotites are characteristic of higher grades, and greenish biotites of lower grades, of regional metamorphism. This point too needs further study.

The distribution of *muscovite* (Fig. 7d) is partly the result of the classification of specimens adopted (Table 4), on the assumption that muscovite is unstable at the highest grade of metamorphism. But the distribution in the lower grade rocks is also affected by hydrothermal activity, so that it is not a suitable index for regional metamorphism.

We must also take into account the effects of stress. Actually all these rocks have recrystallized under conditions of stress, as is indicated by their pronounced foliation and lineation and their strongly oriented fabrics. Some speci-

mens, however, show evidence of crushing at a rather late stage: cataclastic textures (bending, cracking, or peripheral granulation of minerals), or mylonization (very fine-grained textures, commonly with porphyroclasts, or with grade metamorphism, tending toward equilibrium at a lower temperature; plagioclase as sodic as albite is rare, but the process is not far advanced. Granitization, on the other hand, apparently has little effect: in the sillimanite-

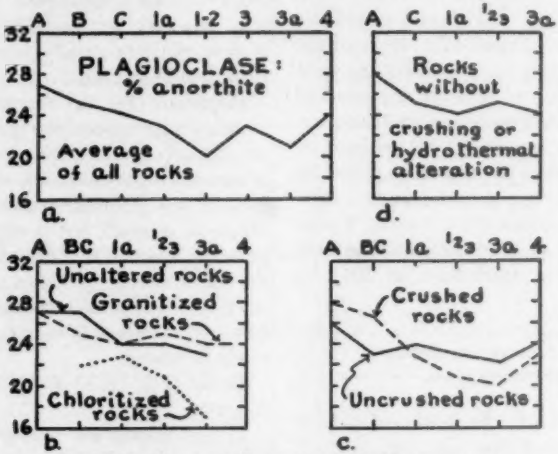

FIGURE 8.—AVERAGE COMPOSITIONS OF PLAGIOCLASES
For groups, see Tables 4, 5, and 6

stringers of oriented and recrystallized quartz). The areal distribution of two textural types is shown on Figure 21, and their distribution by groups in Figure 7h. If the two types be taken together (designated hereafter merely crushed rocks), the graph shows no consistent trend but rather two maxima (at B and at 1–2), indicating that crushing is probably not directly related to intensity of regional metamorphism.

TABLE 5.—REGROUPING OF SPECIMENS TO BRING OUT EFFECTS OF HYDROTHERMAL ALTERATION

	A	BC	1a	1–2–3	3a	4
Number of specimens	23	91	34	131	96	129
Granitized rocks	13	29	5	28	21	127
Chloritized rocks	(2)	32	13	67	34	(2)
Unaltered rocks (no granitization or chloritization)	8	30	16	36	41	0

METAMORPHIC PROCESSES

Effects of Hydrothermal Activity

Study of the effects of granitization and chloritization is facilitated by reclassifying the specimens (Table 5). Since some groups now become very small, groups B and C, and groups 1, 2, and 3 have been combined. Even so, two groups (enclosed in parentheses in Table 5) are so small that they have been omitted in drawing the graphs.

Plagioclase (Fig. 8b) in chloritized rocks is distinctly less calcic than in granitized or unaltered rocks. This is consistent with the idea that chloritization represents a sort of retro-

almandine subfacies, the plagioclase of unaltered rocks is slightly more calcic than that of granitized rocks, while the reverse is true in the staurolite-kyanite subfacies, but the difference is very slight (1–2%) and perhaps not significant.

The antipathetic character of the graphs for red-brown and for yellow-brown *biotites* is striking in unaltered rocks (Fig. 9a, b), but much less so in granitized and chloritized rocks. Yellow-brown biotite is consistently a little higher in chloritized than in granitized rocks (Fig. 9b). Where red-brown and yellow-brown biotites are taken together (Fig. 9c), the trend

already noted (Fig. 7c) is seen to be associated with hydrothermal processes, especially in the staurolite-kyanite subfacies. Figure 9d, e shows the complementary distribution of greenish biotites, green-brown biotite being more characteristic of chloritized, and green-black of granitized, rocks.

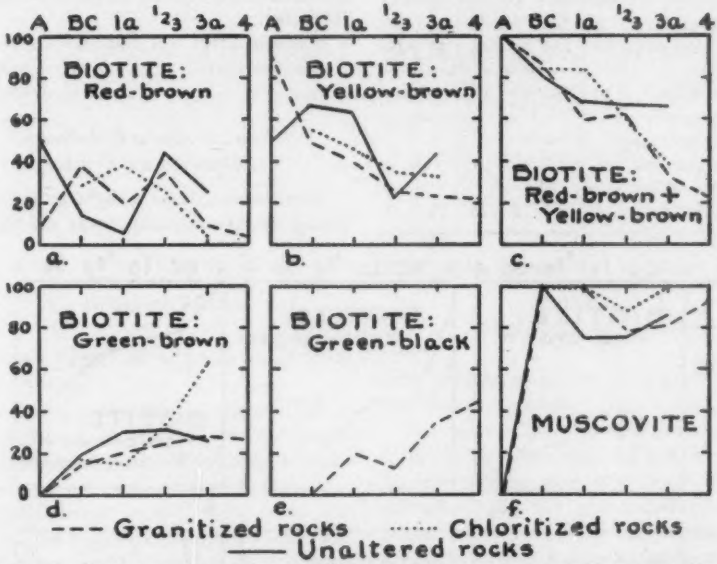

FIGURE 9.—DISTRIBUTION OF MICAS IN GRANITIZED, CHLORITIZED AND UNALTERED ROCKS
For groups, see Table 5.

Rarely does a rock contain more than one type of biotite, and where reddish or yellowish biotites occur with greenish ones the sequence is rarely clear. In a few specimens, red- or yellow-brown biotites show borders of dull-green biotite which may be accompanied by epidote or sphene; the latter is perhaps most significant (Fig. 6): during granitization TiO_2 appears to be released from biotite, recrystallizing in sphene. Whether this is partly controlled by temperature or stress conditions is uncertain. In other cases, borders of greenish biotite are apparently transitional to chlorite. No case has been observed of a red-brown or yellow-brown biotite forming on, or apparently at the expense of, a green biotite.

The shape of the graph for muscovite (Fig. 9f), as already noted, is partly the result of the classification adopted (Table 4)—Group A being defined as containing no muscovite while Groups B and C contain muscovite—but in the other groups the distribution of muscovite is significant. That it is somewhat more common in chloritized than in granitized or unaltered rocks probably indicates that the formation of muscovite continues with falling temperature into the range where chlorite is produced.

Effects of Crushing

Study of the effects of crushing is facilitated by a second reclassification of the specimens (Table 6).

Turner (1948, p. 81) has suggested that the composition of *plagioclase* (Fig. 8c) is to some extent directly dependent on stress, since albite is a stress mineral and anorthite an antistress mineral, and that epidote forms in place of any anorthite in excess of the proportion in equilibrium with the physical environment. This relation may apply to the plagioclase and epidote in the granitized rocks of Group 4, which are thoroughly recrystallized, but it does not seem to apply to the later stages of hydrothermal activity when chloritization was going on. The more sodic plagioclases of the chloritized rocks

TABLE 6.—REGROUPING OF SPECIMENS TO BRING OUT EFFECTS OF CRUSHING

	A	BC	1a	1-2-3	3a	4
Number of specimens	23	91	34	131	96	129
Mylonites	7	35	9	34	14	6
Cataclastic rocks	6	28	8	32	28	23
Total: crushed rocks	13	63	17	66	42	29
Uncrushed rocks	10	28	17	65	54	100
Rocks with no crushing, no granitization, no chloritization	4	7	8	25	24	0

tite more abundant in crushed than in uncrushed rocks. If the two are taken together (Fig. 10c) the influence of yellow-brown biotite predominates, and the rapid decrease in uncrushed rocks of groups 3a and 4 can be correlated with the increase of green-black biotite (Fig. 10e).

Muscovite (Fig. 10f) also shows a tendency, in the staurolite-kyanite subfacies, to be more abundant in crushed than in uncrushed rocks.

Relation between Hydrothermal Alteration and Crushing

Granitization shows a slight tendency to be associated with uncrushed rocks and chloritiza-

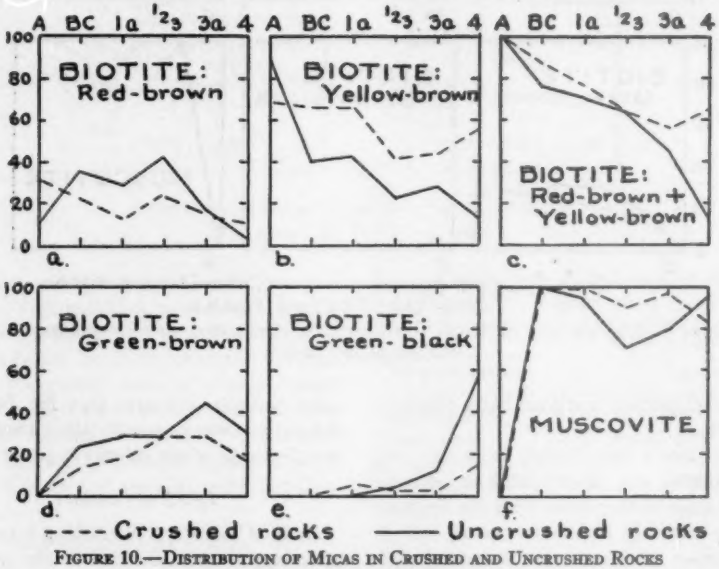

FIGURE 10.—DISTRIBUTION OF MICAS IN CRUSHED AND UNCRUSHED ROCKS
For groups, see Table 6.

are rarely accompanied by epidote or zoisite. It is unexpected to find that the anorthite content is higher in crushed than in uncrushed rocks of the sillimanite-almandine subfacies; this is probably, like some other features of the rocks of this subfacies, a sign of disequilibrium which will be discussed later. In the staurolite-kyanite subfacies, the lower anorthite content in crushed rocks is apparently connected with chloritization.

Biotite (Fig. 10a–e) shows a rather definite correlation with crushing: red-brown biotite is somewhat less abundant and yellow-brown biotion a more marked tendency to be associated with crushed rocks (Fig. 11). This tendency probably reflects differences in the physical conditions governing the two stages of hydrothermal activity. Granitization apparently began at temperatures high enough to permit plastic rather than cataclastic deformation: most of the microcline gneisses are strongly foliated and Ch'ih's (1950) fabric analyses indicate that granitization was, for the most part, syntectonic. But Postel (1940, Pl. 6) has described and illustrated both cataclastic and pseudocataclastic textures in the granitized

rocks. The pseudocataclastic textures record a stage when hydrothermal solutions were penetrating chiefly along channelways formed by minute crush zones, producing fine-grained mosaics of quartz and microcline or myrmekite. The same process, on a larger scale, could account for the "aplite dikes" which locally crosscut the foliation.

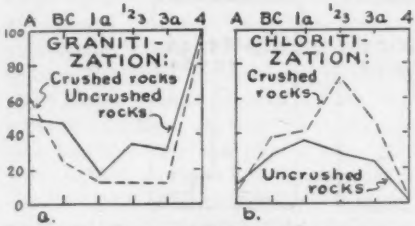

FIGURE 11.—RELATION BETWEEN CRUSHING AND HYDROTHERMAL CHANGES
For groups, see Table 6.

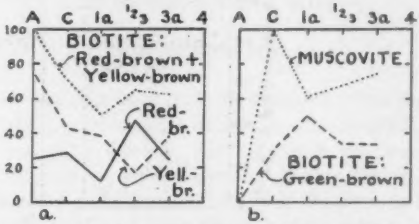

FIGURE 12.—DISTRIBUTION OF MICAS IN ROCKS WITHOUT CRUSHING OR HYDROTHERMAL ALTERATION
For groups, see Table 6.

With falling temperatures, increase of regional stress, or both, ruptural deformation was locally intensified. Postel (1940) mapped crush zones bordering the granitized rocks in the area he studied; similar crushed rocks occur near other, smaller bodies of granitized rocks (Fig. 21). Shearing, with formation of mylonites, may have been localized by the difference in competence between granitized rocks (coarse-grained, composed largely of interlocking quartz and feldspars) and the adjacent schists (largely micaceous). These crushed rocks, in many places, continued to be channelways for hydrothermal solutions which (probably now at a lower temperature) formed stringers of quartz, with muscovite, tourmaline, and chlorite. Similar mineral associations are common along the fault which bounds the schist belt on the north and west.

The distribution of minerals is therefore related to crushing, in so far as crushing facilitated the circulation of hydrothermal solutions, but the effects are somewhat different in rocks of the sillimanite-almandine subfacies and in rocks of the staurolite-kyanite subfacies.

This may be made clearer by a comparison of Figure 11 with Figure 7g, h: In the staurolite-kyanite subfacies, there is little overlapping of granitization and chloritization (Fig. 7g); and granitization is most intense in group 4, where crushing is at a minimum, while chloritization is most intense in group 1–2, where crushing is at a maximum. The geographical distribution is also significant (Figs. 19, 20, 21): chloritization and crushing are largely concentrated in the northwestern half of the schist belt, and granitization in the southeastern half, although some crushed, granitized rocks occur along the northwestern boundary fault.

The rocks of the sillimanite-almandine subfacies, on the other hand, are located mostly in an intermediate zone, with again a few occurrences along the boundary fault. A large proportion of these rocks are crushed (Fig. 7h) and there is considerable overlapping of granitization and chloritization (Fig. 7g): here crushing seems to have favored the penetration not only of chloritizing but also of granitizing solutions (Fig. 11a). Many of these rocks have not attained equilibrium.

Effects of "Pure" Regional Metamorphism

It is evident that hydrothermal activity is responsible for the distribution of most of the minerals so far considered, and that this action has been rendered more effective, especially in later stages, by crushing, locally on a small scale but widespread throughout the region.

Two persistent features, however, distinguish rocks of the sillimanite-almandine subfacies (taken as a whole) from rocks of the staurolite almandine subfacies: (1) somewhat higher anorthite content of the plagioclase, and (2) predominance of red-brown and yellow-brown over greenish biotites. These differences seem attributable to the grade of regional metamorphism, since they persist even when the specimens are reclassified in such a way as to bring

out the effects of hydrothermal alteration or crushing. The same differences appear in the small residue of uncrushed, unaltered rocks shown in Figure 8d, 12a, b.

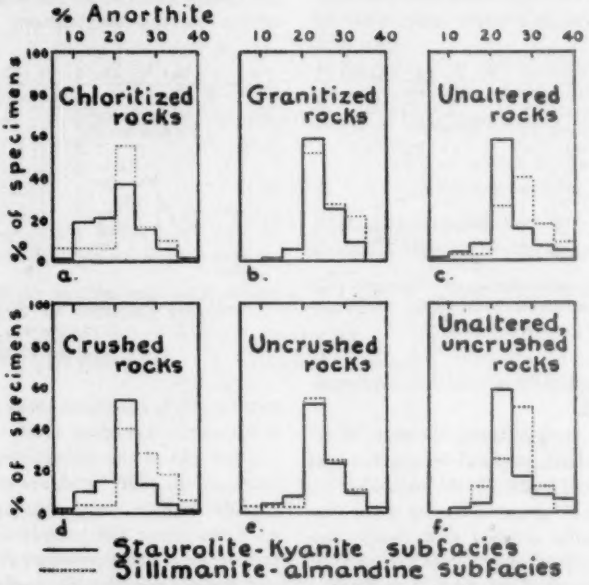

FIGURE 13.—RANGE IN COMPOSITION OF PLAGIOCLASES IN SILLIMANITE-ALMANDINE AND IN STAUROLITE-KYANITE SUBFACIES

Discussion of plagioclase has hitherto been based on Figure 8, which shows the *average composition* (per cent anorthite) for each group. This may usefully be compared with Figure 13, which shows the *range* of composition, in terms of the percentage of specimens in each subfacies (taken as a whole) which fall into classes having anorthite content < 9%, 10–14%, 15–19% etc. The range 20–24% predominates in all the graphs, but chloritization (Fig. 13a) tends, especially in the staurolite-kyanite subfacies, to produce plagioclases with a lower anorthite content, while granitization (Fig. 13b) tends to produce plagioclases with a higher anorthite content. Crushing (Fig. 13d) extends the range in both directions, an effect apparently due chiefly to chloritization in the staurolite-kyanite subfacies and to granitization in the sillimanite-almandine subfacies. In uncrushed rocks (Fig. 13e), the differences between the two subfacies are negligible. The differences are most marked (Figs. 13c, f) where the effects of hydrothermal alteration have been eliminated: plagioclases in the staurolite-kyanite subfacies show a fairly regular distribution about a maximum at composition 20–24% anorthite, and plagioclases in the sillimanite-almandine subfacies are distributed, somewhat less regularly, about a maximum at composition 25–29% anorthite.

These distributions suggest that the plagioclases of the two subfacies formed under slightly different physical conditions, the more calcic ones probably representing somewhat higher temperatures than the more sodic ones. Differences in amount of available CaO may also have existed: if CaO was introduced during granitization, it will have affected the "border zone" adjacent to the granitized rocks where most of the rocks of the sillimanite-almandine subfacies are found. Moreover, departures from the "equilibrium composition" may in some cases represent original differences in composition, as in the rock reported by Ch'ih (1950, p. 934, 948, specimen 12) which has in adjacent layers plagioclases differing 11% in anorthite content (23% and 34%).

The previous discussion of biotite may be

summarized here. Red-brown and yellow-brown biotites apparently contain more TiO_2 than green-black biotites. The relations of sphene and biotite may be comparable to those of epidote and plagioclase: under one set of physical conditions, there is only one phase—titaniferous biotite or calcic plagioclase; under different physical conditions, two phases appear—green biotite + sphene, or sodic plagioclase + epidote. The critical environmental factors certainly include temperature, but in both cases, stress and, perhaps, the chemical character of pore solutions may also play a part. Ch'ih (1950, p. 930) has shown how during granitization the formation of microcline at the expense of plagioclase may produce epidote (zoisite) as a "by-product", and her equations suggest that a change may likewise occur in the relative proportions of (Mg, Fe)O and Al_2O_3 in biotite. The latter change perhaps affects the solubility of TiO_2 in biotite, with sphene forming as a "by-product"; but TiO_2 may also have been redistributed by hydrothermal solutions, or the original content of TiO_2 may have differed in different rocks.

The compositions of plagioclase and biotite probably are, at least in a general way, "indices of regional metamorphism"; but since both must depend also upon the amount of CaO, or of MgO, FeO, and TiO_2 available, it would be unwise to rely upon these minerals alone in determining the metamorphic grade of any individual specimen. On a statistical basis, however (Figs. 8d, 12a) it may be inferred that some of the specimens included in groups 3 and 3a of the staurolite-almandine subfacies actually represent a higher grade of metamorphism, since their composition is such that the assemblage almandine-micas, or micas alone, remains unchanged in intermediate subfacies (Fig. 3b, c).

Finally, the graphs for muscovite (Figs. 9f, 10f, 12b) suggest that some of the muscovite in the staurolite-almandine subfacies has been formed during crushing and hydrothermal alteration, possibly at the expense of other aluminous minerals. In the sillimanite-almandine subfacies also muscovite has been formed, at least in part, by alteration. It is noteworthy that Group B, which on other grounds (see discussion of Table 4) appears to be in disequilibrium, has no specimens free of crushing or hydrothermal alteration, but Group C is represented by uncrushed, unaltered rocks in which muscovite may be stable. These rocks could be assigned to an intermediate sillimanite-muscovite subfacies.

Aluminosilicate Index Minerals

Descriptions of Minerals

The next question to be discussed is how far the formation of the index minerals almandine, staurolite, kyanite, and sillimanite has been affected by hydrothermal processes and by crushing. Petrographic study of the form and associations of these minerals offers some evidence.

Almandine is probably the most stable of these minerals. It occurs throughout the whole temperature range represented by the rocks of this region, not only in unaltered but also in chloritized and granitized rocks, even in the presence of microcline. Barth (1936, p. 820) has assumed that almandine and microcline may occur in equilibrium; but the diagrams I have used (Figs. 1, 4) assume that almandine-microcline is not a stable association. Figure 14a, which shows the per cent of garnetiferous rocks with *euhedral* crystals of almandine offers some evidence for this assumption. Nearly all the euhedral almandine is in unaltered or chloritized rocks, and the rare specimens found in granitized rocks are associated not with microcline but with epidote or zoned plagioclase. Apparently, granitizing solutions tend to corrode and destroy crystals of almandine, probably converting them to biotite; but since granitization was syntectonic, the materials have been redistributed and recrystallized instead of forming recognizable pseudomorphs. Chloritization has had little effect on garnets, except in a few crushed rocks where broken crystals have cracks filled with chlorite.

Figure 14b shows for comparison the distribution of euhedral almandine in crushed and in uncrushed rocks; in the latter, some large crystals have been shattered and drawn out into lenticles, but in the highly micaceous rocks of group 3, although garnets have commonly been rolled (Weiss, 1949, p. 1694, fig. 2), they were protected from breakage by the slipping of surrounding micas.

Staurolite. Turner (1948, p. 82) suggests that the temperature range of staurolite is narrow; probably its chemical range is also limited. It seems to be incompatible with potash feldspars: in the sillimanite-almandine subfacies it occurs

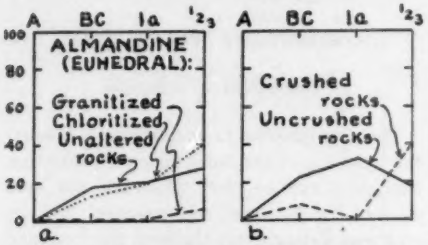

FIGURE 14.—DISTRIBUTION OF EUHEDRAL ALMANDINE CRYSTALS
For groups, see Tables 5 and 6.

TABLE 7.—GROUPING OF SILLIMANITE AND KYANITE ROCKS

	Sillimanite	Sillimanite with kyanite	Kyanite
	S	SK	K
Number of specimens	48	37	58
Granitized rocks	14	12	14
Chloritized rocks	16	13	22
Unaltered rocks	18	11	24
Crushed rocks	32	26	34
Uncrushed rocks	16	11	24

with sillimanite but not with orthoclase; in the staurolite-almandine subfacies with kyanite and almandine, but not with microcline. Euhedral crystals are not common, and broken crystals in crushed rocks are partly replaced by chlorite. No graphs have been prepared for staurolite, since the number of specimens is too small.

Kyanite and sillimanite. The relations of the Al_2SiO_5 minerals are complex. Rocks containing sillimanite and kyanite have been regrouped (Table 7): the group (SK) in which sillimanite is associated with kyanite represents a state of disequilibrium and is presumably transitional between rocks containing only sillimanite (S) and those containing only kyanite (K). Both minerals, however, occur in granitized and chloritized as well as in unaltered rocks, and their distribution does not suggest that the formation of either one at the expense of the other is directly connected with hydrothermal activity.

The *habit* of both minerals, however (Fig. 15), does seem to be influenced by the conditions of recrystallization. Sillimanite occurs in three forms: (1) euhedral prisms—so rare that they are not included in the graphs; (2) compact masses of fibrolite; (3) fine needles or radiating tufts penetrating other minerals—almandine, micas, plagioclase, or quartz. Many specimens contain both fibrolite and needle sillimanite.

Kyanite occurs in two forms: (1) euhedral to subhedral bladed crystals, common in micaceous layers of schists; (2) small stubby needles, in radiating bunches bordering or penetrating plagioclase or myrmekite (Ch'ih, 1950, Pl. 2, figs. 1, 2). These minute needles are difficult to distinguish from sillimanite; in doubtful cases crushed material was mounted in liquid (n = 1.690) for identification. Both needles and blades of kyanite may be found in the same specimen.

The needles of sillimanite and kyanite (Fig. 15a) are somewhat commoner in granitized and chloritized than in unaltered rocks; while fibrolite and bladed kyanite are commoner in crushed than in uncrushed rocks (Fig. 15d).

The common association of needle kyanite with myrmekite and plagioclase suggests that it may be a "by-product" of the granitization process: if soda or potash feldspar is formed at the expense of anorthite, the substitution of NaSi or KSi for CaAl in the feldspar structure might release Al atoms which could recrystallize in kyanite. But the plagioclases associated with needle kyanite show no systematic differences as compared with the plagioclases of other rocks in the same groups; probably, then, the needle kyanite is not simply formed *in situ*, but is the product of more complex reactions in which migrating solutions play a part.

The same remarks may apply to needle sillimanite, at least where it occurs in the interstices between quartz and feldspar grains. It is also commonly associated with mica, from which it may be derived. In many cases, however, petrographic evidence indicates that muscovite has formed at the expense of sillimanite

(and orthoclase?) (Ch'ih, 1950, p. 933, Pl. 2, Fig. 4). Needles (whether sillimanite or kyanite is uncertain) have been completely replaced by muscovite, or sillimanite survives only in the centers of large crystals of almandine or mica,

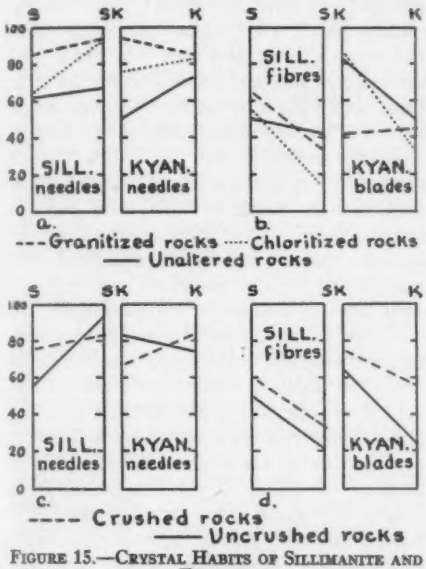

FIGURE 15.—CRYSTAL HABITS OF SILLIMANITE AND KYANITE

For groups, see Table 7.

where it has been protected from reaction with later hydrothermal solutions.

The relation of sillimanite and kyanite to crushing is also complex. Kyanite is commonly considered to be a stress mineral, and its formation seems to have continued to a fairly late stage, since it occurs with micas in crush zones between porphyroclasts of garnet, plagioclase, or micas. Some kyanite crystals are themselves crushed or bent. Sillimanite as knots or radiating swirls of fibrolite appears to be of earlier formation. In some crushed rocks, it occurs only within porphyroclasts of almandine or muscovite; in others, lenticles of fibrolite are more or less intact, but are bordered by grains of kyanite mixed with muscovite.

Fibrolite, like orthoclase and more calcic plagioclases (Fig. 8c), may thus be regarded as a relict, preserved in uncrushed layers or lenticles of crushed rocks. Where recrystallization, aided by hydrothermal solutions, has been complete, sillimanite has been replaced by muscovite or kyanite, orthoclase by muscovite or microcline, and more calcic by less calcic plagioclase.

Mineral Sequences

The inclusion of one mineral by another is a notoriously uncertain criterion for determining the order of crystallization of minerals in metamorphic rocks, but in this case such data seem worth listing for comparison with other evidence.

Almandine is commonly poikiloblastic, whether euhedral or anhedral. Besides quartz, micas, magnetite, and rarely chlorite or tourmaline, some almandines enclose staurolite, kyanite, or sillimanite. This of course does not prove that almandine was the latest mineral to form; it was probably forming at the same time as many of the other minerals, since it has a wide temperature range and a strong "force of crystallization".

Staurolite is also commonly poikiloblastic, including quartz, magnetite, or some minute indeterminate particles concentrated in the cores of crystals. In contact with kyanite, staurolite is generally anhedral, appearing to be moulded on or partially enclosing kyanite. But here, too, the greater "crystallizing force" of kyanite may account for its euhedral form relative to staurolite.

Sillimanite and kyanite occur together in many specimens, but evidence as to their relative age is scanty except in some crushed rocks in which sillimanite is confined to porphyroclasts or uncrushed lenticles, with kyanite and micas in the intervening crush zones. In these the sequence is clear: (1) formation of sillimanite, (2) crushing and formation of kyanite. Weiss (1949, p. 1752) states that "sillimanite in many places appears to be pseudomorphous after kyanite", but she gives no localities and I have been unable to find any such material. Among the local specimens of sillimanite in the collections of Bryn Mawr College, some are radiating or curved masses, flattened and more or less conforming to attached layers of mica schist, but none offers any evidence of being formed directly from blades of kyanite. Of the 85 thin sections of sillimanite rocks used in this study, none con-

tains demonstrable relics of kyanite in process of alteration to sillimanite.

Kyanite also occurs as pseudomorphs after andalusite. Gordon (1922, p. 90–91) lists a number of localities for andalusite, all within one rather narrow belt along Crum Creek

TABLE 8.—FIELD ASSOCIATIONS OF SILLIMANITE, KYANITE, AND PEGMATITES
(Based on Gordon, 1922)

	Sillimanite	Sillimanite with kyanite	Kyanite	"Andalusite" kyanite
Number of localities	23	9	14	11
Associated with				
pegmatite	21	7	12	7
microcline	10	2	4	4
albite (plumasite, corundum)	3	1	1	2

(Fig. 22). The collections of Bryn Mawr College contain 14 specimens of this material: squarish prisms, 1 to 6 inches long, terminated by basal pinakoid; the crystals are grayish, rough or spangled with mica flakes, occurring as radiating groups in a matrix of grayish quartz and mica. Fragments of some of these specimens, studied in immersion liquids, consist mostly of kyanite. Thin sections cut through the centers of two different crystals, each about an inch in diameter, show only a granular mass of small kyanite crystals, with a little quartz, mica, and magnetite. There is no trace of andalusite, even as a core.

The sequence of crystallization suggested by this set of data would be: (1) sillimanite and andalusite (mutual relations unknown), (2) kyanite, (3) staurolite?, (4) almandine?. This evidence is perhaps only negative, in that the sequence is almost the reverse of what might be expected if these index minerals formed consecutively during normal progressive metamorphism.

Field Relations

Pertinent data on the field relations of sillimanite and kyanite are summarized in Table 8, based on Gordon's (1922) lists of mineral localities. This publication, since it brings together the records of many different observers, some of them for exposures no longer accessible, supplies a quite independent line of evidence. Of the 32 localities from which sillimanite has been reported, 28 are described as "pegmatite", and for 12 of these microcline is listed among minerals found; 4 are albite (plumasite) dikes with corundum, or associated with hydrothermally altered ultrabasic rocks. There seems little doubt that sillimanite (at least as large specimens of interest to mineral collectors) is almost invariably associated with pegmatites. Similar associations are common also at localities where kyanite and "andalusite" (kyanite pseudomorphs) have been found. Staurolite and garnet are less commonly reported as associated with pegmatites, and the garnet in such cases is often listed as spessartite, not almandine.

Finally, the areal distribution of these minerals may be reconsidered. Comparison of Figures 18 and 19 shows that sillimanite is most abundant in or near areas of granitization. In a few places, notably at the northeastern and southwestern ends of the schist belt, sillimanite is adjacent to bodies of gabbroic rocks. It is difficult to ignore the conclusion that sillimanite has been formed in connection with either igneous intrusion or regional granitization, though this is emphatically denied by Weiss (1949, p. 1703–1704).

Kyanite to a large extent shares the distribution of sillimanite (Figs. 17, 18), but it is also found, like staurolite and almandine, well outside the granitized areas.

SUMMARY AND CONCLUSIONS

Subfacies and Isograds

The region contains rocks of several different "grades" of metamorphism, evidently formed under different physical conditions. The highest "grade" is characterized by sillimanite or orthoclase; the next by kyanite, staurolite, muscovite, or microcline; the lowest by chlorite. The validity of these three divisions is supported by differences in the plagioclases and biotites: in the highest "grade" the average anorthite content of plagioclase is over 25 per cent, and the predominant biotite is yellowish- or reddish-brown (probably containing TiO_2); in the next, the anorthite content of plagioclase averages under 25 per cent, and greenish

biotites (in some rocks accompanied by sphene) are common; in the lowest, plagioclase may be still more sodic (to 5 per cent anorthite), and biotite is partially replaced by chlorite. These three "grades" may be correlated with the subfacies designated by Turner (1948) the sillimanite-almandine, staurolite-kyanite, and albite-epidote amphibolite subfacies.

The changes representing the lowest grade of metamorphism are of a retrograde character and are only incipient. None of the rocks so affected has reached equilibrium in the albite-epidote subfacies, and it would be impossible to draw a corresponding isograd in this region, although rocks of this subfacies are found farther to the northwest.

The other two subfacies would theoretically be effectively separated by a "sillimanite isograd". Each of the two subfacies may again be subdivided if we assume that the transitional groups already discussed, kyanite-almandine (1a) and sillimanite-muscovite (B, C), are valid subfacies representing intermediate conditions of metamorphism.

Isograds drawn in accordance with this assumption (Fig. 22) show an interesting and fairly simple pattern, perhaps indicating a locus of high grade metamorphism in the southwestern end of the schist belt, with regularly declining zones to the northeast, north, and northwest. But this simplicity is in some ways deceptive; for when this map is compared with Figures 17 and 18, irregularities appear which demand further discussion.

Chemical Composition

The 22 chemical analyses discussed earlier (Tables 1, 2, 3; Figs. 4, 5) may now be reconsidered in relation to metamorphic zones.

The staurolite zone is represented by analyses D, 1, 2, 3, 4: Of these, D and 3 correspond to rocks containing staurolite; the others represent rocks containing micas or garnets, presumably because their composition did not favor the production of staurolite.

The kyanite-almandine zone is represented by analysis C. Specimens from this locality contain microcline and almandine, but not kyanite; the original composition of the rock seems to have been altered by granitization. Specimens from locality B contain kyanite and almandine, but if the isograds are correctly drawn, this locality is within the next zone. Overlapping of zones will be discussed later.

The sillimanite-muscovite zone is represented by analyses A, B, 5 and 12: Of these, 12 corresponds to a specimen containing sillimanite and muscovite; B, as already noted, is from a locality with kyanite and almandine; and A and 5 from localities with almandine or micas only.

The sillimanite-almandine zone is represented by analyses J, 10, 11, 13; analysis 13 has the mineral assemblage characteristic of this zone. The others correspond to specimens having kyanite or muscovite with orthoclase.

Analyses E, F, G, H, 6, 7, 8, 9 all contain microcline and represent granitized rocks.

The influences of chemical composition upon the mineral assemblages produced during metamorphism has two aspects:

(1) The original composition of the rock may or may not permit the formation of the index minerals characteristic of each zone, even when equilibrium has been attained. In the staurolite zone, many rocks contain almandine or micas only; these, however, are at the same "grade" as the accompanying staurolite rocks. I believe that there is no "garnet zone" in this region; staurolite reappears on the northwestern flank of the Baltimore gneiss block, and chloritoid rocks occur along the same line of strike farther west. The staurolite isograd, marking the outermost edge of the staurolite-almandine subfacies, lies in the general position indicated by McKinstry (1949, p. 882).

(2) The same explanation may well account for rocks containing almandine or micas only which lie within the kyanite-almandine or sillimanite-muscovite zones. But the possible effects of granitization cannot be excluded. On Figure 4, analyses A, C and 5 all show a rather higher proportion of K_2O than most of the other schists, and the last two are close to areas of granitized rocks. Here too the chemical composition evidently precludes the formation of sillimanite or kyanite, but it is at least questionable whether this was the original composition of the sediments.

With still higher proportions of K_2O (Fig. 4, analyses E-H, 6-9), the rocks become microcline gneisses or "granodiorites". Such rocks are found in all four zones. Their original character has undoubtedly been altered by

granitization during a period of hydrothermal activity which accompanied or followed the most intense phase of metamorphism in the region.

Overlapping of Zones

The transition from one "grade" of metamorphism to the next cannot, of course, be instantaneous, and a certain amount of overlapping in the distribution of index minerals is to be expected. In this region, however, the persistence of many "lower grade" minerals into "higher grade" zones is probably the result of a complex metamorphic history.

The kyanite-almandine isograd has been drawn along the outer edge of the area where rocks containing this mineral assemblage occur; on the scale of these maps, this line closely approximates a "kyanite isograd". If the assemblage kyanite-almandine replaces assemblages with staurolite at slightly higher temperatures, this isograd ought to be closely followed by one marking the disappearance of staurolite. This is not the case: rocks with staurolite persist not only into this, but also into "higher" zones (notably south of localities 7 and 12. Compare Fig. 22 with Fig. 17). Moreover, some of the sillimanite rocks contain staurolite (Table 4).

Similarly, the sillimanite-muscovite isograd fails entirely to separate rocks with kyanite from those with sillimanite. Kyanite rocks are found sporadically throughout the whole sillimanite-muscovite zone, and many sillimanite rocks contain kyanite as well (Table 4).

The sillimanite-almandine isograd is the most deceptive of all. Only about 15 per cent of the rocks within this isograd really belong to this subfacies: most of them contain muscovite; many contain kyanite. Also, this isograd roughly encloses the largest body of "granodiorite" in the region—rocks containing muscovite, microcline, myrmekite, and epidote, which certainly formed at temperatures lower than those assumed for the sillimanite-almandine subfacies.

One general conclusion seems to be justified: each increase in the "grade" of metamorphism is accompanied not only by the appearance of "higher grade" mineral assemblages, but also by an increase in the number of "lower grade" assemblages which seem to be out of equilibrium.

Retrograde Metamorphism

In one sense, any region showing metamorphic zoning represents progressive metamorphism, since the maximum temperature could not have been reached instantaneously; but the rocks of the highest grade need not have passed successively through all the stages of equilibrium represented by lower grades, if the increase of temperature were fairly rapid, as for instance where igneous intrusions supply heat.

The stage of declining temperatures, on the other hand, is presumably a long one, at least in deep-seated metamorphism. The fact that retrograde metamorphism is not more frequently observed is generally attributed to the decrease in the velocity of many chemical reactions as temperatures fall, and to the tendency of many high-temperature minerals to persist, as either stable or unstable relicts, even under changed physical conditions. In the Philadelphia region, however, the stage of declining temperatures was accompanied by events which facilitated retrograde metamorphism-tectonic movements continuing to a late stage of crushing and mylonization, and introduction of copious hydrothermal solutions. Thus the regional pattern of metamorphism is a palimpsest, on which the earlier traces of progressive metamorphism have been strongly over-written with a record of retrogressive changes.

In this sense, the sillimanite-almandine isograd shown on the map (Fig. 22) is a relict feature. It has already been stated that most of the rocks within this isograd have minerals characteristic of lower zones. Petrographic evidence indicates the direction of the change: partial replacement of orthoclase by microcline, myrmekite, or muscovite, and replacement of sillimanite by muscovite and (at least in crush zones) by kyanite. At a more advanced stage of this replacement, rocks with the assemblages sillimanite- (almandine-) muscovite or kyanite-almandine (-muscovite) would be formed, and these are in fact common in this zone. The final stage is the production of microcline gneiss or "granodiorite", which occupies a large part of the zone.

This interpretation raises the question of the origin of the sillimanite-muscovite and kyanite-almandine rocks in adjacent zones. Billings

(1950) maintains that the assemblage sillimanite-muscovite is stable and constitutes a separate subfacies. This may be true, but it is difficult to prove in the Philadelphia region. Many of the rocks in this zone show partial replacement of sillimanite by muscovite and many contain kyanite; and signs of granitization are common, though no large masses of "granodiorite" are found. In my opinion, a majority of these rocks have been produced by granitization of rocks which once contained a larger quantity of sillimanite and probably orthoclase—i.e., they are "retrograde" representatives of the sillimanite-almandine subfacies. The status of the sillimanite-muscovite subfacies will probably have to be further tested in regions where hydrothermal activity is minor or absent.

Discussion of the kyanite-almandine zone and its relation to higher zones involves some discussion of the more general problem of the relation between sillimanite and kyanite. Much has been written on this subject which need not be reviewed here. The hypothesis which best fits the facts in this region is one recently put forward (Turner and Verhoogen, 1951, p. 412) that sillimanite is the stable form of Al_2SiO_5, kyanite and andalusite appearing only under certain favorable conditions. Tilley (1935) has also suggested that kyanite forms as a metastable intermediate phase in some cases. It is not then the persistence of sillimanite in retrograde metamorphism that demands an explanation, but the formation of kyanite.

Unfortunately the conditions favoring the development of kyanite are not well understood. Stress is often stated to be a controlling factor, and this may be responsible, in this region, for the conversion of andalusite to kyanite, and indeed perhaps for most of the kyanite in the kyanite-almandine zone. In many of the granitized rocks, however, Al_2SiO_5 has evidently been redistributed by hydrothermal solutions, recrystallizing as fine tufts or needles of sillimanite or kyanite in interstices between grains of quartz and plagioclase or in myrmekite. Sillimanite or kyanite pegmatites apparently represent the same process on a larger scale. For these occurrences, I would venture the hypothesis that the chemical character of the circulating solutions—perhaps their content of alkalis or of "acid" hyperfusibles such as HF or HCl—may determine the form in which Al_2SiO_5 crystallizes.

The existence of staurolite in association with sillimanite can be explained by the failure of kyanite to form under certain physical or chemical conditions.

In the kyanite-almandine and staurolite zones, granitization is of minor importance—pegmatites and a few small bodies of "granodiorite", locally accompanied by sillimanite. The dominant type of retrograde metamorphism is that already described as chloritization: there are two substages, one of tourmaline formation and one of muscovite and chlorite formation, but the two overlap. Tourmaline probably indicates the higher temperature stage, for it occurs in many pegmatites and some granitized rocks without chlorite; in crushed rocks, broken or "stretched" tourmalines have cracks filled with chlorite. Chemical change was slight: The soda content of plagioclase has been slightly increased, and biotite has been partially chloritized, but almandine, staurolite and kyanite have been little affected.

The relation between granitization and chloritization is schematically shown in Figure 16, which should be considered in connection with Figures 19 and 20. The chief center of granitizing solutions was along the southeastern margin of the schist belt; here temperatures were higher than elsewhere, solutions more copious and chemically more active. Reconstitution and recrystallization of rocks was locally nearly complete, with the formation of "granodiorites" characterized by microcline and green biotite with sphene. Myrmekite and sodic rims on plagioclase, together with muscovite symplectite and epidote, probably continued to form at slightly lower temperatures and are found in rocks farther out from the main center of granitization. Toward the north and west, the intensity of hydrothermal activity decreased and the original composition of the sediments was not so much altered. As temperatures throughout the whole region declined still further, a certain amount of cataclastic deformation continued (Fig. 21). In the granitized belt, myrmekite, muscovite, and epidote, with some tourmaline, were still forming; while farther out, tourmaline and chlorite were the characteristic minerals of the last phase.

Although the general relations are clear, it

is impossible to draw precise "isograds" for this retrograde metamorphism, since the amount and effectiveness of hydrothermal solutions were controlled in part by the permeability of different rocks and the localization of crush zones.

thus took place not at the maximum, but at a somewhat lower, temperature; it was prolonged, so that large areas of rocks were completely granitized ("granodiorites"), and rocks of the sillimanite-almandine subfacies suffered diaphthoresis. This stage may be said to have ended

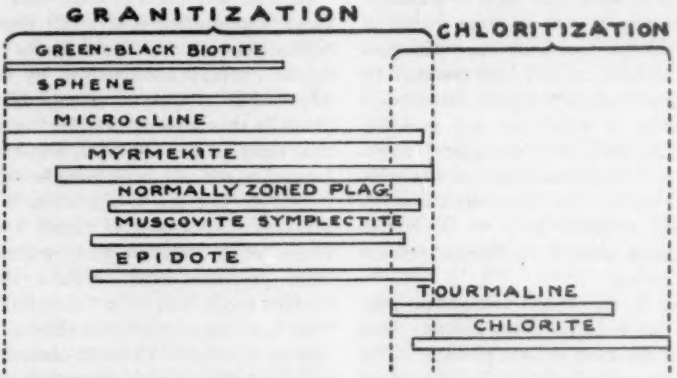

FIGURE 16.—RELATION BETWEEN GRANITIZATION AND CHLORITIZATION

Metamorphic History of the Region

An attempt may now be made to reconstruct the metamorphic history of the region:

1. The earliest stage that can be inferred is probably connected with the intrusion into the Wissahickon sediments of basic and ultrabasic igneous rocks, now best preserved along the western margin of the schist belt. Stress was not intense, and conditions were perhaps those of contact rather than regional metamorphism: sillimanite and andalusite were formed, and there may even have been a rather extensive zone of andalusite hornfelses.

2. The next stage was marked by the introduction of granitic material, probably as hydrothermal solutions. At the same time, regional deformation was initiated or intensified. In the southeastern half of the schist belt, rocks of the sillimanite-almandine subfacies were formed; farther northward, temperatures were those of the staurolite-kyanite subfacies and granitizing solutions were less active except in a few patches. Andalusite was converted to kyanite, and strongly foliated schists were formed.

These conditions continued for some time, though with temperatures slowly declining and regional stress perhaps intensified. This is the main stage of regional metamorphism, which at temperatures characteristic of the staurolite-kyanite subfacies. Microcline-muscovite rocks were stable, orthoclase and sillimanite persisting only as relict minerals.

3. As temperatures throughout the region declined still further and deformation entered the ruptural stage, hydrothermal solutions produced muscovite, tourmaline, and chlorite. These changes are more conspicuous in the northwestern half of the schist belt where temperatures were probably always lower and hydrothermal solutions less active than in the southeastern zone of granitization, but the total effect is slight.

The dating of this metamorphism can be only conjectural. As all who have studied the problems of Piedmont geology are aware, the Wissahickon formation itself has been variously assigned to the Precambrian, the Cambrian, and the Ordovician; this controversy need not be discussed here.

The three stages of metamorphism outlined above may be separated by considerable intervals of time, and may even belong to different geologic periods ranging from the end of Precambrian to the end of Paleozoic time.

On the other hand, they may be parts of a continuous process connected with one igneous cycle, which began with intrusions of basic

magmas, continued with orogenic movements and the emplacement of granitic material (to which the residual liquids of the basic magmas may have made some contribution), and finally waned with a period of cataclastic deformation and low temperature hydrothermal activity.

If the latter interpretation is accepted, the metamorphism would probably be dated as Paleozoic. The northern boundary fault which separates the Wissahickon formation from lower Paleozoic rocks truncates the latter; and rocks along this fault (Rand, 1901; Armstrong, 1941) locally contain porphyroblasts of muscovite, epidote, and abundant tourmaline, indicating that the crushed rocks of this zone were serving as channelways for hydrothermal solutions similar to those active elsewhere in the schist belt. The Paleozoic rocks are also cut by a few small pegmatites and locally have abundant tourmaline. If this mineralization belongs to the sequence described in this paper, the period of metamorphism can be no earlier than Ordovician, and more likely is to be correlated with the Appalachian revolution.

Note on Maps, Figs. 17–22

The base for the maps, Figs. 17–22, was drawn from parts of U. S. Geological Survey quadrangles: Norristown, Pa.; Germantown, Pa.; Burlington, Pa.–N. J.; Chester, Pa.–Del.–N. J.; and Philadelphia, Pa.–N. J.

Geologic boundaries have been taken mostly from U. S. Geological Survey Folios 162 (Philadelphia) and 167 (Trenton); for granitized areas west of Philadelphia, from Postel (1940). Outlines have been generalized and many small bodies omitted entirely on account of the small scale of the maps.

Data plotted on the separate maps have been obtained chiefly from the study of 504 thin sections, including materials used in the preparation of the Philadelphia and Trenton Folios (Bascom, 1909a,b) and specimens collected by several other workers cited (Postel, 1940; Postel and Adelhelm, 1944; Cloos and Hietanen, 1941; Weiss, 1949; Ch'ih, 1950).

Additional information on the distribution of sillimanite, kyanite and staurolite (Figs. 17, 18) was obtained from the following sources: study of rock fragments, crushed in a mortar and mounted in immersion liquids; study of heavy mineral separations mounted in balsam, including materials collected by Bell (1940), Weiss (1949), and Dike (1950); and Gordon's (1922) list of mineral localities. Where information is lacking on associated minerals such as muscovite and orthoclase, special symbols have been used for sillimanite and kyanite.

The distribution of orthoclase and microcline (Figs. 18, 19) was also checked by a study of 25 additional samples of rock fragments mounted in balsam.

Numbers and letters indicating groups on Figures 17 and 18 correspond to those of Table 4a and b.

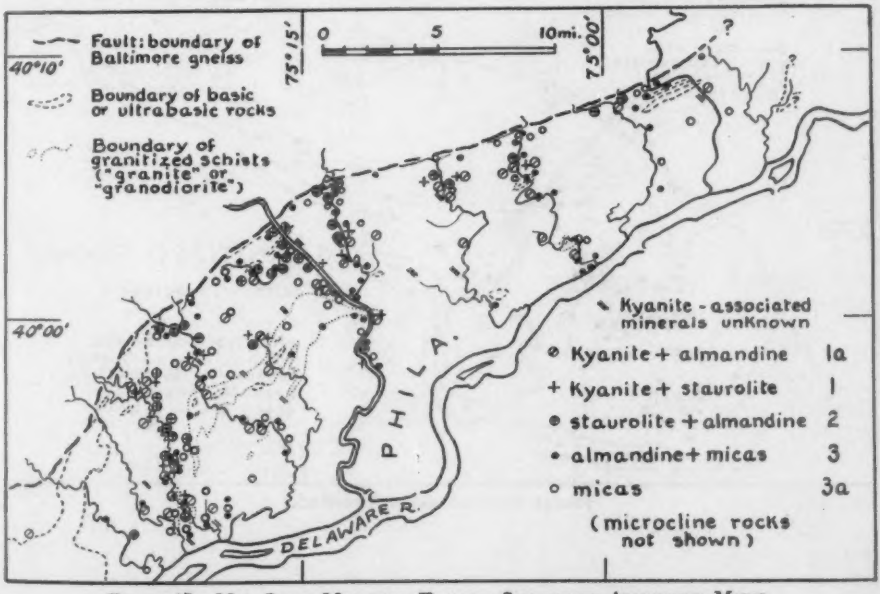

Figure 17.—Map: Index Minerals: Kyanite, Staurolite, Almandine, Micas

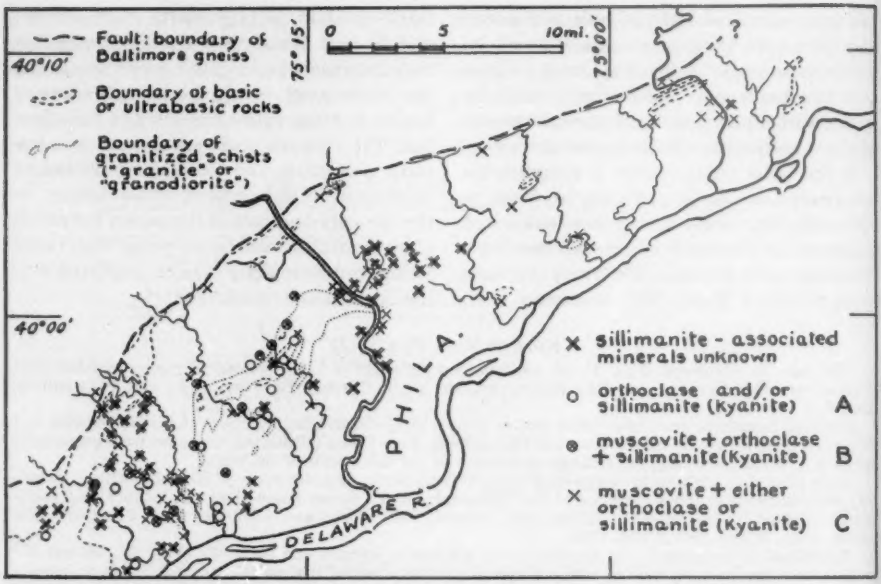

FIGURE 18.—MAP: INDEX MINERALS: SILLIMANITE (KYANITE), ORTHOCLASE, MUSCOVITE

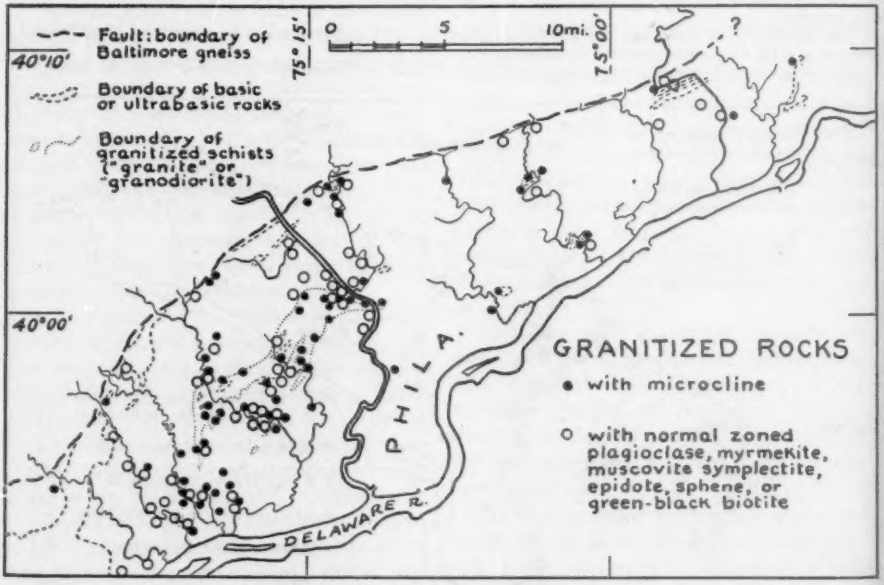

FIGURE 19.—MAP: GRANITIZED ROCKS

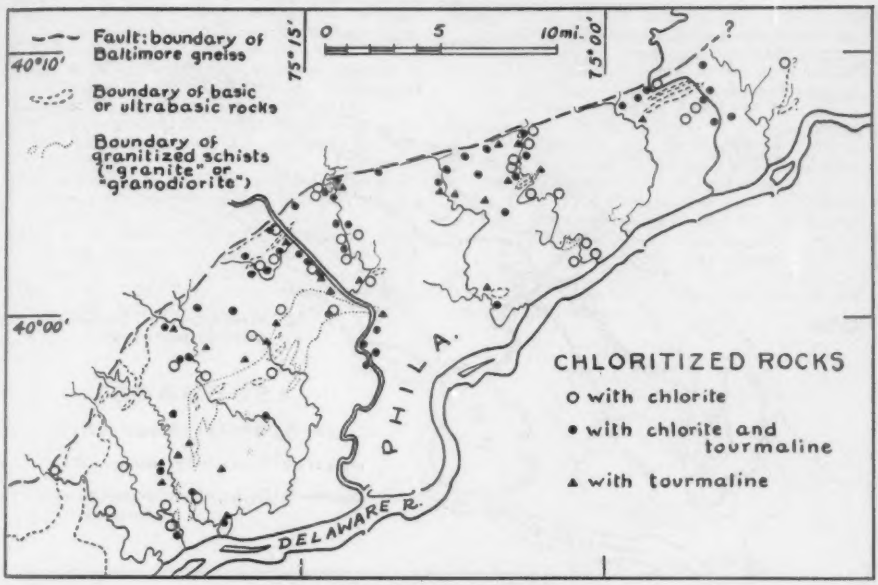

FIGURE 20.—MAP: CHLORITIZED ROCKS

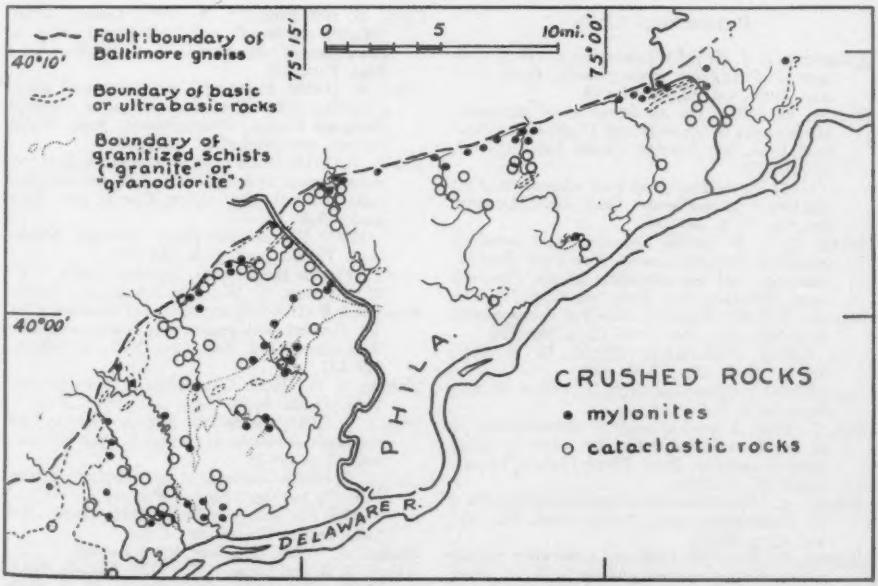

FIGURE 21.—MAP: CRUSHED ROCKS

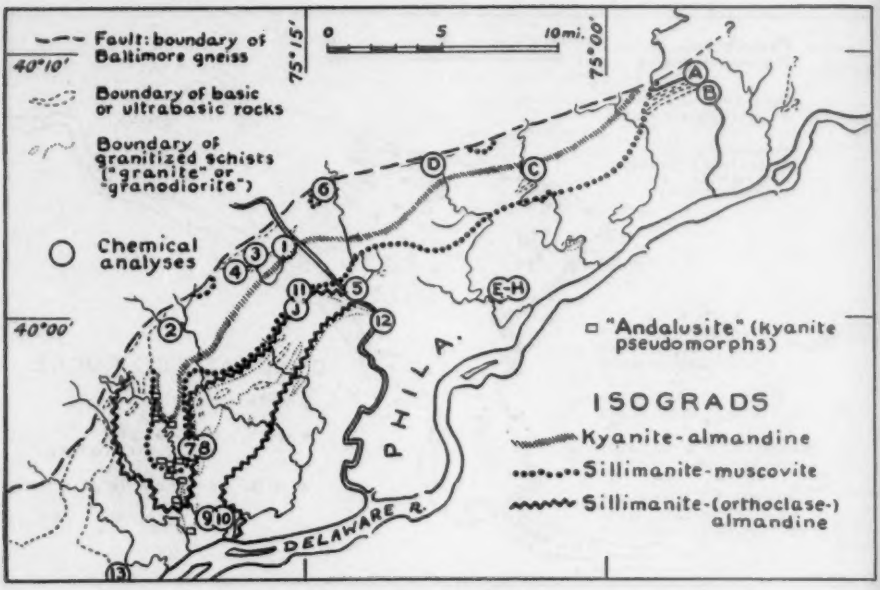

FIGURE 22.—MAP: ISOGRADS

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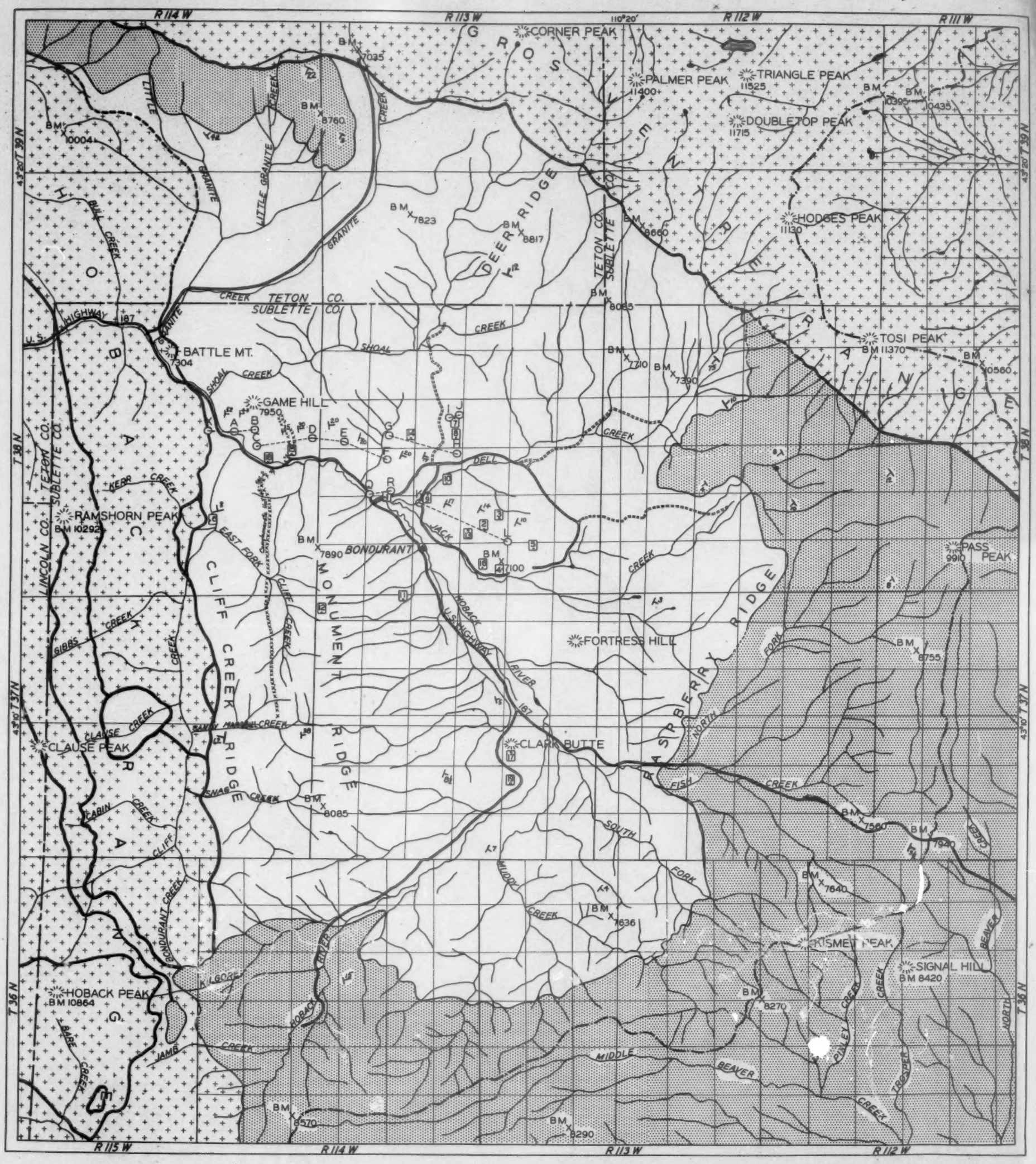

GEOLOGIC MAP OF HOBACK BASIN AND VICINITY, WYOMING
(Modified from Eardley, 1944)

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EARLY CENOZOIC STRATIGRAPHY AND VERTEBRATE PALEONTOLOGY OF THE HOBACK BASIN, WYOMING*

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ABSTRACT

Several episodes of Laramide deformation occurred in ranges bordering the Hoback Basin. Deformation, beginning in the west, provided sediment for the Hoback formation. The Hoback formation was overriden by post-Wasatchian thrusting. The middle Eocene? Pass Peak conglomerate overlapped the thrusts; was involved in later faulting. Interpretations of the complex regional structure are reviewed.

The Hoback formation is over 15,000 feet thick. A type locality for the formation is designated and the beds there are described. Other sections are combined with the type section in a composite columnar section. Fossil localities are listed. The coal, carbonaceous shale, and abundant plant fragments; drab coloration of the beds; and irregular bedding, indicate fluviatile deposition in a forested, humid, floodplain environment. An early Tiffanian arboreal and forest mammalian faunule (from the Dell Creek Quarry) 9000 + feet above the base of the formation supports this conclusion. The lower two-thirds of the formation are Paleocene in age.

Vertebrates from the Hoback formation are described. Their significance is discussed. New species are described for each of the mammalian genera: *Ptilodus?*, *Plesiadapis*, *Carpodaptes*, *Didymictis*, *Haplaletes*.

Coryphodon and *Probathyopsis successor* are reported from a higher locality. Early development of sexual dimorphism in the Uintatheriidae is recognized.

An early Wasatchian mammalian faunule is described from a still higher horizon.

The Hoback formation is partly equivalent in age to the Almy formation but resembles Paleocene and earliest Eocene deposits in the Jackson Hole Area lithologically. Deposition probably began during early Paleocene time, continuing without major interruption into Wasatchian time.

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* Modified and condensed from a dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy in the University of Michigan.

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Introduction

Location and General Description of Area

The Hoback Basin (Fall River Basin of Schultz, 1914) covers an area of approximately 250 square miles in the northwestern corner of Sublette County, and small adjoining portions of Lincoln and Teton counties, in central western Wyoming (Fig. 1). The basin lies directly north of the Green River Basin and, although not separated by an intervening mountain range, the two belong to entirely different drainage systems. The Green River Basin is drained by the Green River which flows southward as the main tributary of the Colorado River. The Hoback Basin is drained by the Hoback River which arises in the southern part of the Hoback Range, flows northeastward into the basin, and then swings around in a broad arc toward the northwest. It cuts almost directly across the Hoback Range in a deep canyon about 20 miles north of its source and flows into the Snake River 12 miles due south of Jackson, Wyoming. The Hoback Basin, with a mean altitude of less than 7000 feet, is at least 500 feet lower than the Green River Basin. It is roughly triangular in outline. Its northern apex is formed by the convergence of the Hoback and Gros Ventre ranges in the vicinity of Little Granite Creek. It terminates on the west against the north-south trending Hoback Range which rises to an altitude of nearly 11,000 feet at its southern end. The northeastern margin is formed by the abrupt face of the Gros Ventre Range which trends NW and includes peaks over 11,000 feet in height. Toward the southeast the basin opens broadly, separated from the Green River Basin by the Hoback Basin Rim, the low drainage divide between the Green and Hoback rivers. This divide rises to an altitude of 9910 feet at Pass Peak, but elsewhere is somewhat lower. U. S. Highway No. 187, which traverses the basin, is the main route between Rock Springs and Jackson, Wyoming, and crosses the divide at an altitude of 7941 feet.

Bedrock is poorly exposed in the basin. Although numerous sandstone horizons outcrop, these have yet to yield any fossils. The shale horizons are poorly exposed and these exposures are usually of limited extent. No true badlands exist within the basin. This makes the discovery of fossils difficult. Most of the area is accessible by dirt road or trail though some is almost inaccessible. The geography of the area is well covered in earlier reports (St. John, 1883, p. 202–206; Schultz, 1914, p. 13–27).

Field and Laboratory Work

Field work on this problem was begun during the summer of 1947; only one specimen was acquired during the short intervals spent in the area. During the summer of 1948, the writer again spent short periods of time and a University of Michigan paleontological field party under the direction of Dr. Claude W. Hibbard and accompanied by George R. L. Gaughran and Walter H. Wheeler spent 2 weeks, in the area. Several specimens were recovered from the Dell Creek Quarry locality at that time, and Dr. Hibbard found a molar tooth of *Hyracotherium* at Locality No.

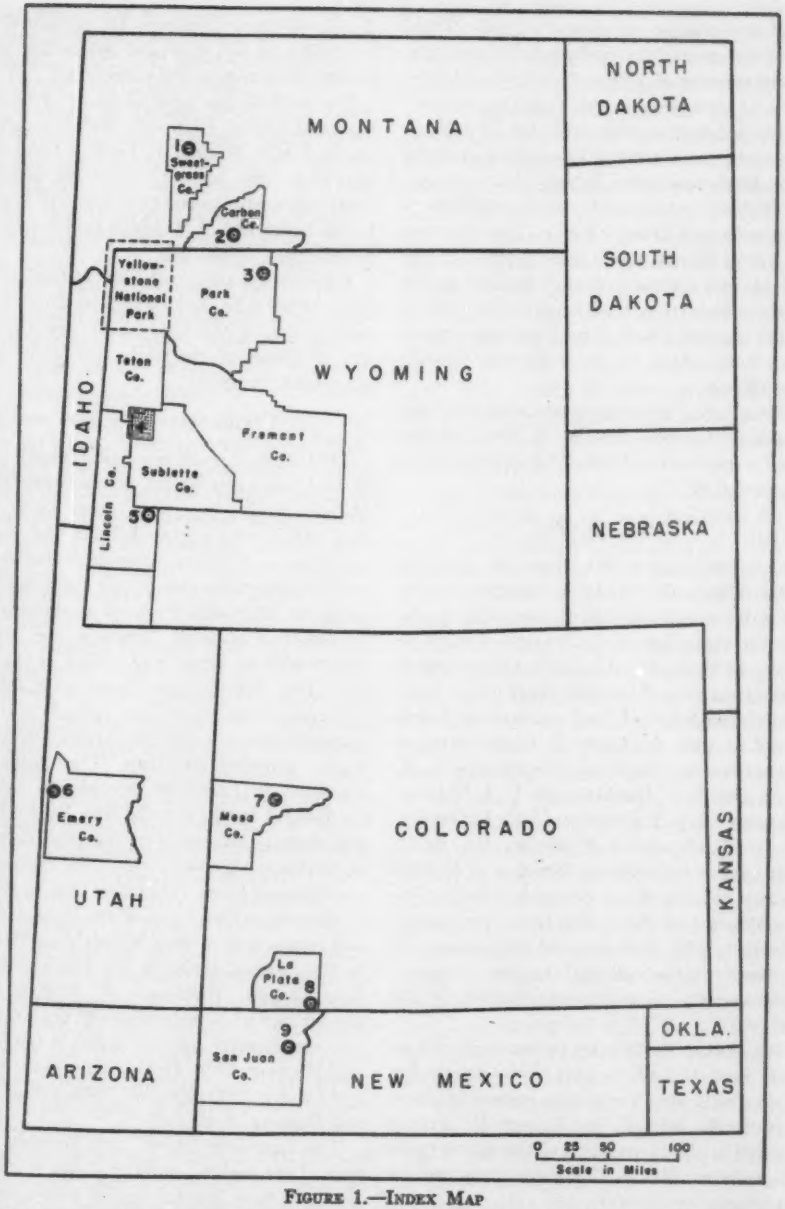


FIGURE 1.—INDEX MAP

Showing location of area of study and principal Paleocene mammal localities. (1)—Crazy Mountain field local faunas, (2)—Bear Creek local fauna, (3)—Polecat Bench local faunas, (4)—Hoback Basin, area of this study, (5)—Labarge Creek local fauna, (6)—Dragon and Wagonroad local faunas, (7)—Plateau Valley local faunas, (8)—Tiffany local fauna, (9)—Puerco and Torrejon faunas.

2 and invertebrates at several other localities. The writer spent 3 full months in the area during the summer of 1950 collecting fossil vertebrates at previously known localities, measuring detailed stratigraphic sections in the Hoback formation, and searching for additional vertebrate fossil localities. During that summer, Dr. Hibbard and his party, which consisted of Thomas Oelrich, Dwight Taylor, Tom Sparrow, and James Rogers, again spent 2 weeks in the area. On two occasions, student geology groups from the University of Michigan Camp Davis visited the Dell Creek Quarry and some specimens were added to the collection through their efforts.

Almost all of the preparatory work has been done by the author. Mr. W. H. Buettner prepared a specimen of *Plesiadapis* and the uintathere material.

Acknowledgments

In the field and in the laboratory, the help of Dr. Claude W. Hibbard, chairman of the writer's doctoral committee, has been of the greatest value. Dr. A. J. Eardley of the University of Utah offered valuable suggestions as to the treatment of the field phase of the work. Direct comparison of the fossil material herein described with specimens in their respective institutions was made possible by Drs. C. L. Gazin and D. H. Dunkle of the U. S. National Museum, Dr. J. T. Gregory of Yale University, Dr. G. L. Jepsen of Princeton, Dr. Bryan Patterson of the Chicago Museum of Natural History, and Dr. G. G. Simpson of the American Museum of Natural History. Dr. Jepsen offered helpful discussion of the taxonomic problems involved although he does not necessarily endorse the conclusions reached by the author.

Mr. Walter R. Wheeler examined the uintathere material and the part of the manuscript dealing with that group. His suggestions were particularly helpful. Dr. Robert V. Kesling assisted in the preparation of the photographs of specimens. Mr. W. C. Sherman prepared the drawings of specimens.

The original manuscript, from which this paper is taken, was critically read by Professors W. H. Burt, E. N. Goddard, R. C. Hussey, L. B. Kellum, and K. K. Landes of the University of Michigan. Dr. Hussey visited the writer in the field and personally added several specimens to the collection.

Residents in the area of study were most kind and helpful. Particular thanks are due to Mr. and Mrs. William T. Faris of Bondurant, Wyoming, who not only aided the author in many ways, but opened their home to him and to the University of Michigan field parties as a headquarters in the field.

Without the continual help and encouragement of my wife, Ruth Pritchett Dorr, in the preparation of the manuscript and throughout all phases of the work, this study would have been impossible.

General Geology

Pre-Cenozoic rocks are well exposed in the Hoback and Gros Ventre ranges bordering the Hoback Basin; in certain other areas, they crop out in very narrow belts at the margins of the basin. Nowhere, however, do outcrops of pre-Cenozoic rocks extend any appreciable distance into the basin. Rocks of all systems from Precambrian through Mesozoic age are included with the possible exception of the Silurian. Two distinct depositional facies, foreland and geosynclinal, have been brought into close juxtaposition by Laramide thrusting from two nearly opposing directions. The variance in character of Laramide deformation between the Hoback Range and the Gros Ventre Range was strongly influenced by the great difference in thickness of these facies. The thickness of pre-Cenozoic strata varies from over 15,000 feet in the geosynclinal facies of the Hoback Range and ranges farther west, to less than 5500 feet in the foreland facies of the Teton and Gros Ventre ranges (Wanless, *et al.*, 1946, p. 1240). Recent partial considerations of the pre-Cenozoic stratigraphy may be found in the following: Horberg, 1938; Dobrovolny, 1940; Nelson and Church, 1943; Eardley, 1944; Foster, 1947; and Horberg, *et al.*, 1949.

The Laramide history and geologic structure of the region are too complex to be fully reviewed here. A great deal of recent work by geology students and staff members of the University of Michigan Camp Davis bears directly upon the problem, but has not yet appeared in print. Hence, only those structural

aspects which have a direct bearing upon the relationship between the older rocks bordering the Hoback Basin and early Cenozoic formations within it will be mentioned.

The beds of the Hoback formation dip generally eastward, their steepness increasing toward the west. At the junction of Cliff and Kerr Creeks on the western side of the basin, however, they strike N80°E-dip 8°W, and along Sandy Marshall Creek, immediately east of the Game Thrust (of Eardley, 1944) strike N6°E-dip 62°W. They have been folded by Laramide compressional forces beneath the easternmost (Game Thrust of Eardley, 1944) of a series of several imbricate low angle thrusts. This overthrust relationship is clearly evident at Battle Mountain near the mouth of Granite Creek where the salmon-colored lower Jurassic Nugget sandstone and reddish-brown lower Triassic shales of the Woodside and Thaynes formations overlie beds of the Paleocene Hoback formation above a nearly horizontal contact. This easternmost thrust can be traced southward from Battle Mountain nearly to Kilgore Creek where the trace of the fault passes beneath a series of younger Cenozoic beds referred to as the Pass Peak conglomerate (Eardley *et al.*, 1944). The trace of the thrust is undulant and is marked for the most part by the strongly outcropping Nugget sandstone. North of Battle Mountain, along the western side of Granite Creek, the fault trace is obscured by alluvium and glacial debris and its position is in dispute. Eardley (1944) shows it to be continuous with the Game Thrust, thus forming part of a continuous thrust traceable southeastward from near the town of Jackson almost to Little Granite Creek and thence southward nearly to Kilgore Creek where it passes beneath the Pass Peak conglomerate. Horberg *et al.* (1949, Pls. 1, 2; p. 201–202) and Nelson and Church (1943, p. 161–162) do not recognize the Game Thrust as being continuous with the easternmost thrust. Their concept is more complex: The thrust along the greater part of the western margin of the Hoback Basin is called the Cliff Creek Thrust and is shown as terminating a short distance north of Battle Mountain. Farther north two unnamed arcuate thrust slices are shown as doubtfully related to the Little Granite Thrust. Horberg *et al.* consider the Little Granite Thrust of later origin, overriding the Pass Peak conglomerate about midway between Horse Creek and Little Granite Creek. Horberg *et al.* imply that some of the thrusting toward the east, involving the geosynclinal facies, took place after the deposition of the Pass Peak conglomerate. Eardley's interpretation implies that this thrusting preceded the deposition of the Pass Peak conglomerate. The lithologic character of the beds at the northwestern corner of the Hoback Basin, west of Granite Creek, and mapped by both authors as being the Pass Peak conglomerate, is distinctly different from that of the conglomerates at Pass Peak. It has been suggested (Horberg *et al.*, 1949, p. 190) that these beds may represent a marginal facies of the Hoback formation. No paleontologic evidence can yet be applied to this question.

West of the easternmost thrust in the southern Hoback Range, at least five other thrusts have been recognized (Eardley *et al.*, 1944). It is not known whether these thrusts become progressively older or younger toward the west. They are probably of nearly the same age; that is, thrusting took place after the deposition of the Hoback formation and before that of the Pass Peak conglomerate. Farther south, near the head of the south fork of the Hoback River, Eardley (1944) has shown the Lookout Thrust. Its relationship to the Pass Peak conglomerate is not clear. Unfortunately, Eardley's structural cross-sections do not cross this area. The outcrop of the Pass Peak conglomerate is shown as terminating against the trace of the Lookout Thrust. Whether this thrust overrides the Pass Peak or whether the Pass Peak simply reaches but does not overlap the thrust is not certain from the literature and no evidence can be contributed here. Horberg *et al.* (1949, p. 211), from their study of Eardley's (1944) map, believe the former to be the case and cite this as evidence, in conjunction with their belief in similar relationships at the northwestern margin of the basin, for post-Pass Peak thrusting in the Hoback Range. The writer has followed Eardley (1944) for the eastern boundary structure of the Hoback Range in preparing the map (Pl. 1). The Game Thrust between Battle Mountain and Kilgore Creek is equivalent to the Cliff Creek Thrust of Horberg *et al.* Where

the positions of these faults are doubtful, they are represented by broken lines.

Along the northeastern margin of the basin, the structural relationships are in even greater dispute. Horberg *et al.* (1949, *see* Plates 1 and 2; p. 195–198) and Nelson and Church (1943, *see* Fig. 6–8; p. 150–198) recognize thrust faulting (the Cache Thrust) at the northwestern end of the range, but terminate this thrust against the high-angle reverse Skyline Trail Fault between Horse and Little Granite creeks. Southeast of there, they show the essential structure as a series of three high-angle reverse "trapdoor" faults. Lying basinward from these faults are two wedges of steeply southwestward-dipping Carboniferous to Jurassic rocks which they represent as upfaulted in part against the Hoback formation, but these supposed subsidiary faults are shown as dying out toward the southeast so that in part the pre-Cenozoic strata are overlapped by the Hoback formation. Horberg *et al.* (1949, p. 198), however, consider it possible that these faults may be part of a continuous thrust (the Cache Thrust). Eardley (1944) has mapped the Cache Thrust as being continuous. Along the western side of Granite Creek, conglomerates supposedly of the same age (but, as already pointed out, of different lithologic characteristics) as the Pass Peak conglomerate have been dragged upward where they rest against older downdragged Paleozoic rocks along the front of the Gros Ventre Range. This strongly suggests considerable post-Pass Peak movement along the marginal fault structure at least at this point.

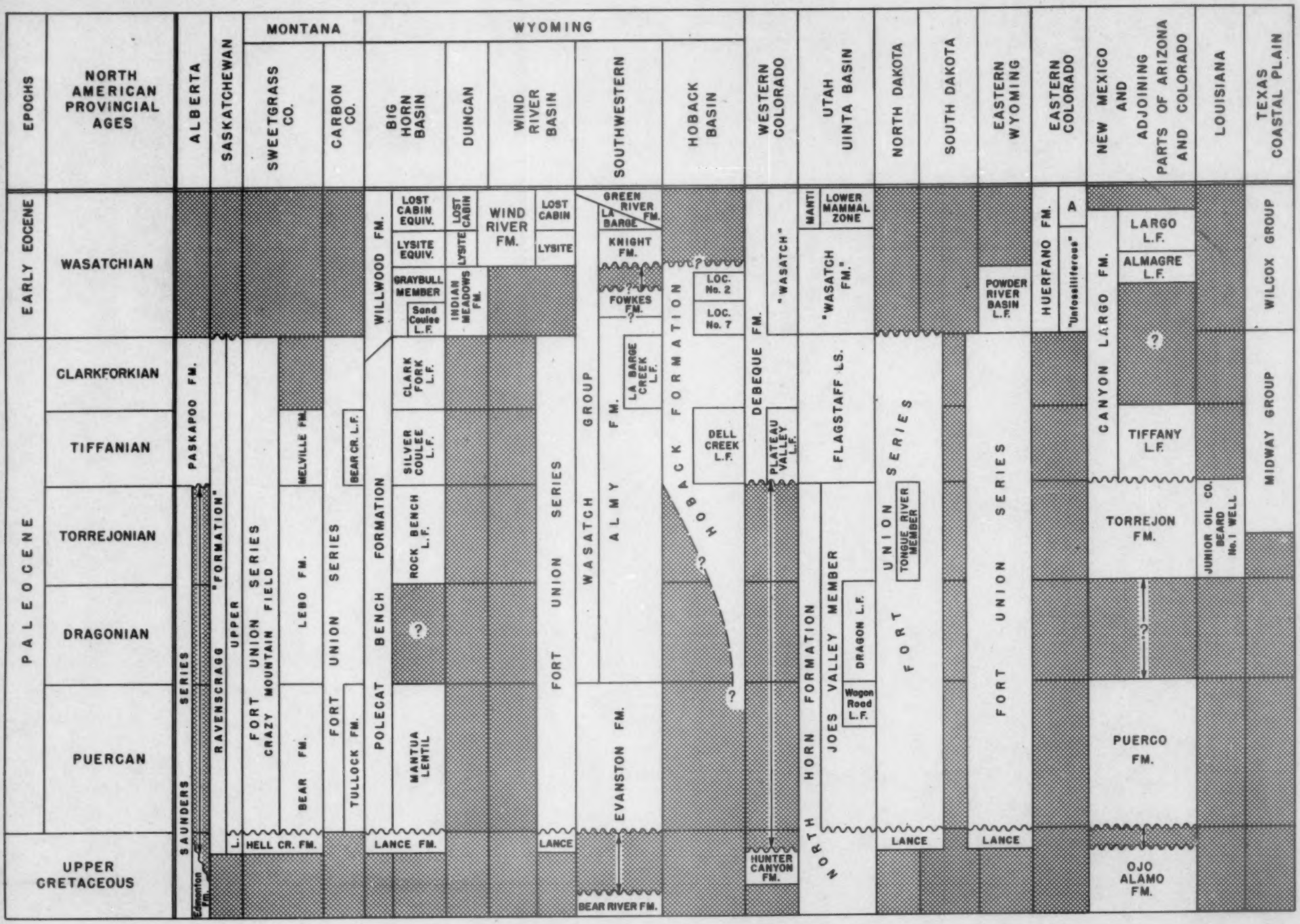
A full appreciation of the intimate relationship between marginal Laramide structures and the Cenozoic beds of the basin is essential since dating of these beds will ultimately result in a more certain understanding of the sequence of regional Laramide deformation. In summary, on the western margin of the basin thrusts toward the east have overridden the Hoback formation and have subsequently been overlapped by the Pass Peak conglomerate. Some evidence has been advanced that certain of these eastward thrusts may also have overridden the Pass Peak conglomerate. Thus, some of the geosynclinal deformation took place in post-Hoback pre-Pass Peak time, and some possibly in post-Pass Peak time. Along the northeastern margin of the basin, the faulting may have taken place either before, or after, the deposition of the Pass Peak conglomerate or possibly both before and after. Another possibility is revealed by the view of Horberg *et al.* that major faulting in the Gros Ventre Range and folding and some thrusting in the Hoback Range may have occurred prior to the deposition of the Hoback formation.

Early Cenozoic Deposits

Hoback Formation

Distribution.—The early Cenozoic beds, herein referred to as the Hoback formation, crop out so far as is known only in the Hoback Basin and are particularly well exposed for about 4 miles southeastward from Battle Mountain along U. S. Highway No. 187 on the northeastern side of the Hoback River. The unshaded central part of the map (Pl. 1) shows the exposed limits of the formation. It passes beneath the Game Thrust in the Hoback Range along the western margin of the basin. On the northeast, it terminates at the foot of the Gros Ventre Range as described earlier. Toward the southeast, it is unconformably overlain by the Pass Peak conglomerate. Excluding local covering by glacial deposits and alluvium its outcrop area comprises about 200 square miles.

Previous studies.—The first geologist to describe the Cenozoic beds of the basin was Orestes St. John, a member of a government survey party which traversed the basin in 1878. In his report (St. John, 1883) the basin beds are referred to in most places simply as "Tertiary" and are undifferentiated, although in one sketch (St. John, 1883, Pl. 14) the area corresponding approximately to that in which the Hoback formation crops out is labelled "Tertiary-Wasatch" and in the vicinity of Pass Peak the beds are labelled "Tertiary, red conglomerate." In general his report (Pls. 9, 13–15) represents "Tertiary" beds as resting nonconformably upon strata ranging from Silurian to late Cretaceous in age. The bottom cross-section on Plate 9, however, shows them normally down-faulted against the "Carboniferous and Triassic" along the front of the Hoback Range in the vicinity of the South Fork of the Hoback River. St. John (1883, p. 204) de-



CORRELATION CHART OF NORTH AMERICAN PALEOCENE AND EARLY EOCENE
(Modified from Wood *et al.*, 1941)

scribed beds along the east flank of the "Hoback Canyon Ridge" (Hoback Range) and no doubt belonging to the Hoback formation as:

"light grayish, slightly calcareous clays and shales [which] incline off the monoclinal ridge of steeply dipping, non-conformable Jura-Trias at angles of inclination varying from 15°–45° eastward, gradually flattening out toward the center of the basin."

Blackwelder (1911, Pl. 17; Fig. 53, p. 465) shows the location of the Fall (Hoback) River and Fall River Basin and represents the Cenozoic beds along the eastern flank of the Hoback Range in the vicinity of Hoback Canyon as normally downfaulted against older strata, but gives no lithologic description or indication of their age.

Schultz (1914) considered the Cenozoic deposits of the "Fall River Basin" in some detail. He stated that: "The thickness of the [Evanston] formation in the Fall River Basin is at least 9500 feet; how much more is not known for the top of the beds was not seen. (p. 69)" On the same page, he states:

"The Evanston formation in the Fall River Basin reaches a maximum thickness of about 9500 feet. The beds generally dip 15°–45° E., but farther west, toward the older beds, they dip more steeply. The Evanston in this basin lies unconformably upon Weber quartzite and later Carboniferous, Triassic, and Jurassic beds, involving an unconformity of more than 20,000 feet. The rocks consist chiefly of clay, carbonaceous shale, shaly sandstone, and brown, gray, yellow, and white sandstone. Associated with these rocks are some coal beds, but owing to the extensive talus cover, to their ready weathering, and to lack of prospecting it is not possible to state how thick or numerous they are. Probably they contain as much coal as the Adaville formation east of Labarge Ridge."

Schultz (p. 77) suggested that the principal Laramide structures in this region were formed before deposition of the Evanston and Almy (Pass Peak conglomerate) formations. At the time he wrote, overthrusting had not been recognized in the Hoback Range.

On the U. S. Geological Survey Geologic Map of Wyoming (Campbell, 1925), an area corresponding roughly to that in which the Hoback formation is exposed is designated "Tertiary Fort Union" and is shown as being overlain by "Tertiary Wasatch." No fault is shown along the eastern front of the Hoback Range on that map.

The U. S. Geological Survey Geologic Map of the United States (Stose and Ljundstedt, 1932) shows the area of outcrop of the Hoback formation as "Eocene Fort Union" and the formation as overlain by "Eocene Wasatch." A fault between the Cenozoic deposits and older formations of the Hoback Range is questionably indicated by a broken line.

Since 1943, the new name, Hoback formation, has been applied to these beds by University of Michigan field parties because evidence for correlation of these beds with the Evanston formation as described by Veatch (1907, Pl. 3; p. 77) was considered too meager. The term was first used in print by Eardley *et al.* (1944, stratigraphic column) who characterized the beds as: "Interbedded gray sandstones and shales. Several conglomerate lenses and freshwater impure limestone beds. Coal in lower part." The thickness was given as 15,000± feet. The formation was considered lower Eocene in age.

Eardley and White (1947, p. 987) employ the formational name, Hoback, in the following connection:

"Again in the Hoback Range and Hoback Basin of Northwestern Wyoming the geosynclinal flysch of the first Laramide disturbance is not found in the immediate hinterland. The molasse is the Hoback formation, some 15,000 feet of shales, arkosic sandstones, and a few conglomerate layers. The molasse is lower Eocene, but the next oldest sediments are those of the Frontier of middle Upper Cretaceous age. A long time separates the two during which the flysch should have been accumulating. The Hoback was overthrust by Mesozoic and Paleozoic strata, and a new molasse was deposited, the Pass Peak conglomerate of middle Eocene age. The setting is actually more complicated than implied in the foregoing description because in nearby regions several formations of upper most Cretaceous and Paleocene age exist, and the spatial as well as the time relations are complex."

Love (1947), in a Tertiary correlation chart, shows the "Hoback formation of the University of Michigan" as late Paleocene (Tiffanian and Clarkforkian) in age, and states that: "The green and brown sandstones and shales, and possibly a part of the underlying Pinyon conglomerate [in the Jackson Hole area], probably correlate with the thick Hoback formation (2) [ref. to Eardley, 1944] on the south side of the Gros Ventre Range."

A U. S. Geological Survey preliminary map by Brown (1949) shows the outcrop of Paleocene beds in the Hoback Basin essentially as on

previous maps, but indicates that Paleocene sediments are coextensive, across the Hoback Basin Rim, with deposits of the same age farther south along the western side of Sublette County. This is considered incorrect since the upper part of the Hoback formation, as shown later, includes sediments as young as early Wasatchian and since (Pl. 1) the overlying middle Eocene (or younger) Pass Peak conglomerate extends across the entire southeastern margin of the basin. The conglomerates and freshwater limestone which flank the western side of the Hoback Range were mapped by Brown (1949) as Paleocene, but are referred to elsewhere (Eardley *et al.*, 1944; Love, 1947) as the Camp Davis formation, and have yielded an upper molar tooth (University of Michigan Museum of Paleontology No. 23494) identified as belonging to the fossil horse, *Pliohippus*, a genus ranging from lower to middle Pliocene (Eardley 1942, p. 1800).

Horberg *et al.* (1949, Table I, p. 187) assign a "Lower Eocene (?)-Paleocene" age to the Hoback formation and indicate its thickness as 5000± feet. They describe the Hoback formation lithologically as: "Interbedded gray sandstone and shale with fresh-water limestone beds and some conglomerate. Fossil invertebrates occur in the limestone. Fossil carnivore jaw found in Hoback Basin."

An accompanying stratigraphic summary (Horberg *et al.*, 1949, p. 186, 190) states:

> "On the basis of fresh-water fossils, the Hoback formation was assigned a lower Eocene age and the Pass Peak formation, which overlies it unconformably, a middle Eocene age. However, an early carnivore jaw found in 1947 is probably late Paleocene in age (A. J. Eardley, personal communication, 1948) and indicates that at least the lower half of the Hoback formation is Paleocene. The deposits are found in both overlap and fault relations with the older formations in the ranges, but the relative extent of these relations is uncertain."

The "carnivore jaw" is the specimen, U.M. M.P. No. 24315, found by the author in 1947 at the Dell Creek Quarry, UM-Sub-Wy Locality No. 1, and identified by Dr. G. G. Simpson as *Thryptacodon australis*, an arctocyonid creodont. The occurrence and significance of this specimen was reported by the author before the Michigan Academy of Science, Arts, and Letters in 1948, but the report was not published. A description of the specimen will be found in the taxonomic part of this paper. The chronology of Laramide diastrophic and depositional phases according to Horberg *et al.* (1949, p. 210–211) may be summarized as follows:

1. "Main folding and faulting in the Gros Ventre and Teton ranges and possibly early folding in the Hoback and Snake River ranges."
2. "Main folding accompanied by some thrust faulting in the Snake River and Hoback ranges."
3. Deposition of the Hoback formation.
4. "Main thrust faulting along the east front of the Hoback Range."
5. Deposition of the Pass Peak conglomerate.
6. Later thrusting in the Hoback Range and possibly in the Snake River Range.

General description.—The base of the Hoback formation is not exposed. The lowest exposed beds within the formation are those immediately east of the Game Thrust along the western margin of the basin. The top of the formation is probably not exposed either, since the Hoback formation is overlain nonconformably by the Pass Peak conglomerate. The exposed part of the formation consists mostly of a monotonous alternation of sandstone and shale. A few impure, argillaceous limestone layers occur, but their combined thickness is insignificant compared with the total thickness of the formation, which is over 15,000 feet. Several carbonaceous shale beds occur in the lower and middle part of the formation. Two miles above the mouth of Little Granite Creek, a lense of sub-bituminous coal of limited extent, with many shale partings, and ranging from a few inches to 3 feet thick, supplies a small amount of coal for local consumption. Near the junction of the Dell Creek road and U. S. Highway No. 187, a small shaft, now abandoned, was driven into a carbonaceous shale horizon (Unit No. 4 of Section No. 5, see Pl. 3) with little apparent success. Schultz (1914, Fig. 7, p. 98) recorded a 3-foot coal seam in the "Fall River (Evanston) coal fields" in the NW¼ of the SE¼ of Sec. 31, T38N, R113W, Sublette County, Wyoming. Coal is not at present being mined from this seam. Coal is not sufficiently widespread in the formation to be of much commercial importance. Fragments of plants, however, are common in many of the sandstones.

The sandstones show some lithologic variability, but are generally light- to dark-gray or brown, weathering to gray, buff, tan, or brown. They range from fine- to coarse-grained, and are arkosic. The grains are angular to sub-angular. The most common constituent is quartz; lesser amounts of feldspar, mica, dark rock fragments, and calcareous cement make up the remainder of the rock. Some of the sandstones are strongly calcareous; others are non-calcareous. Some are massive and well exposed; others are thin, ripple-marked, and irregularly bedded with numerous silty partings. The latter type are poorly exposed or concealed. The bedding is less well developed toward the east where the dip decreases, but not enough to account for the reduction in the number of outcrops of thick sandstones. Part of the explanation lies in the relative increase in shale in the upper part of the formation.

The shales are generally dark gray to dark brown and in a few places black due to the presence of carbonaceous material. Some are calcareous, others are not. Most are non-laminated. Their weathered colors are drab shades of gray and tan, commonly lighter than the unweathered rock.

Two lensing conglomerate horizons, in places reaching considerable thickness, occur in the upper part of the lower third of the formation (Pl. 3, Section 2). Sub-angular to rounded cobblestones up to 12 inches long are common in these units. At least three-quarters of them are quartzites or strongly cemented quartzitic sandstones, although possibly reworked representatives of all local pre-Cenozoic formations are present.

Source of sediment and environment of deposition.—Identification of the source of sediment for the Hoback formation would go far toward clarifying the diastrophic history of the region. The consistent angularity of the grains in the sandstone suggests a local source of sediment. There are several Paleozoic and Mesozoic sandstones in the region which might have served as a source of quartz, particularly such thick upper Cretaceous arkoses as those of the Aspen formation. The angularity of the quartz grains and arkosic nature of most of the Hoback sandstones, however, seem to be evidence that reworking of these older sedimentary formations did not supply the major source of sediment. Precambrian rocks are not exposed in the ranges immediately west of the basin because of the thickness of the geosynclinal facies there and the type of deformation. Hence a Precambrian source of fresh quartz is not to be sought immediately to the west. Some reworked Cambrian and post-Cambrian rocks recognized in the lower conglomerates of the Hoback formation might have been supplied from that direction. In the Gros Ventre Range, the Paleozoic and Mesozoic strata of the foreland facies are thinner and the deformation in that range involved mainly vertical uplift, hence the Precambrian may have been exposed there shortly after uplift began. Thus, it is possible, though not proven, that uplift in the Gros Ventre Range provided a partial source of sediment for the Hoback formation. If the main Laramide folding and faulting in the Gros Ventre Range occurred prior to the deposition of the Hoback formation (Horberg *et al.*, 1949, p. 210–211), the plausibility would be increased. Detailed sedimentary petrologic studies of the Hoback formation are needed before a reliable decision can be reached.

More can be said with regard to the environment of deposition of the Hoback formation. The frequency of lithologic variation in the deposits; the cross-bedding or lack of bedding and the variable thicknesses of the beds; the occasional gravel and conglomerate lenses; the "dirty" appearance of the sandstones; all are characteristics of fluviatile deposition. The majority of the materials were dumped by rivers charged with sediment from the bordering uplands, upon valley flood plains in a basin of low relief at or near sea level. Basin subsidence probably kept pace with deposition. Some of the deposits, such as the limestones with their abundance of molluscs and fish remains, the laminated shales, and coals, accumulated in shortlived sloughs, oxbow lakes, and swamps. The abundance of disseminated plant material and occasional coal seams suggest that, for the most part, the deposits accumulated in a heavily forested area with a warm, humid climate. The universally drab coloration of the formation and the occurrence of an arboreal and forest mammalian faunule in a "pocket" at one

horizon (the Dell Creek Quarry) support this suggestion (Van Houten, 1948, p. 2102–03).

The occurrence of uintathere and coryphodont remains at a higher horizon indicates that either at various times, or in certain places, a savannah environment, with alternating open and forested areas, and probably with the heaviest forest growth along the stream courses, may have developed.

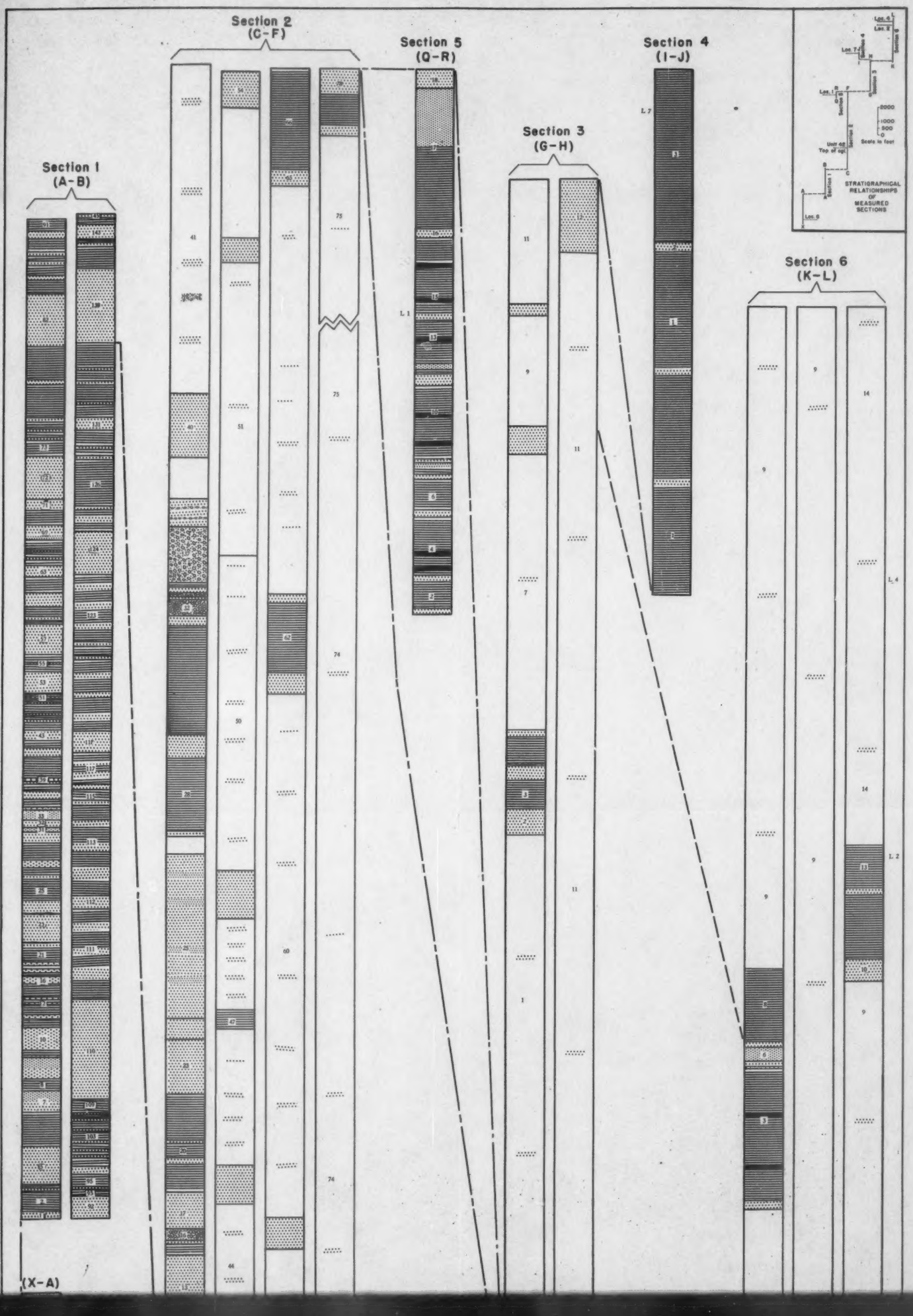
It is impossible to be certain of the rate of deposition of the Hoback formation. In all probability its rate of accumulation was highly variable. The total span of time required for its deposition was longer than has been supposed, however, since the probable age of the formation ranges from at least middle Paleocene up into the lower Eocene. The coarse, thick conglomerates along the present western margin of the basin might be interpreted as marginal facies, but because of the extent of overthrusting one cannot be certain of the original western limit of the basin. It is also possible that the conglomerates reflect accelerated uplift at the source area of the sediment.

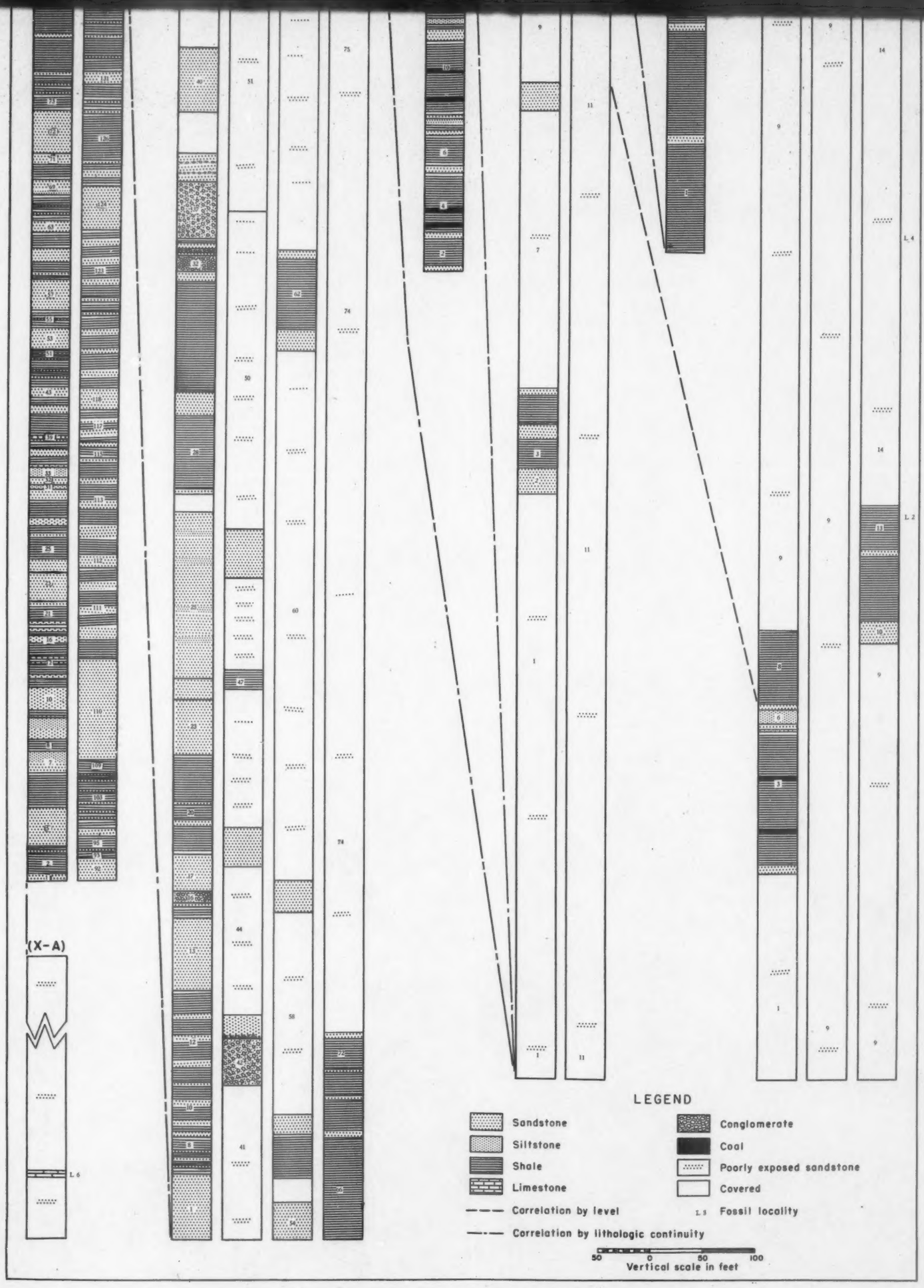
Age and correlation.—The beds of the Hoback formation have hitherto been correlated with the Evanston formation, but there is no good evidence for this. There is no lithologic continuity and only a general lithologic resemblance between the two formations; they occupy distinctly separate basins of deposition. Furthermore, they may not be even partially equivalent in age. Wood *et al.* (1941, Pl. 1; p. 19) tentatively assigned the Evanston formation a Puercan provincial age. Paleontologic evidence is presented in the faunal portion of the present study to prove that approximately the upper third of the Hoback formation includes beds ranging from Tiffanian to at least lowermost Wasatchian in age. Paleontologic evidence does not exclude the possibility that the lower two-thirds of the formation may include beds as old as Puercan. Thus far there is no evidence with which to date the lowermost exposed beds. Certainly the thickness of the formation below the horizon known to be Tiffanian suggests a considerably greater age for the lowermost exposed portion of the formation, below that level. On the correlation chart (Pl. 2), deposition of the Hoback formation is tentatively considered as having begun during the early Paleocene and extended into Wasatchian.

Wood *et al.* (1941, Pl. 1; p. 14, 20) very tentatively considered the Almy formation of southwestern Wyoming as being middle Paleocene in age and the Fowkes formation as upper Paleocene. Gazin, however, (1942, p. 217) has reported an upper Paleocene (Clarkforkian?) mammalian faunule from beds "well down in" the Almy formation, along Labarge Creek in Lincoln County, Wyoming, that were "mapped by A. R. Schultz as the northward equivalent of the Almy formation and are so regarded by Rubey . . ." He further reports the occurrence, at another locality higher in the Almy formation, of *Eohippus* and *Coryphodon*. Hence, it now appears that the Hoback formation may be at least partly equivalent in age to the Almy formation. They are dissimilar in other respects; the Hoback formation is much thicker and rarely shows the reddish tints mentioned by Veatch (1907, Pl. 3; p. 89) in describing the Almy formation. The two formations cannot be shown to be stratigraphically continuous. They probably accumulated in separate basins, under slightly different environmental conditions, and possibly with a different source of sediment.

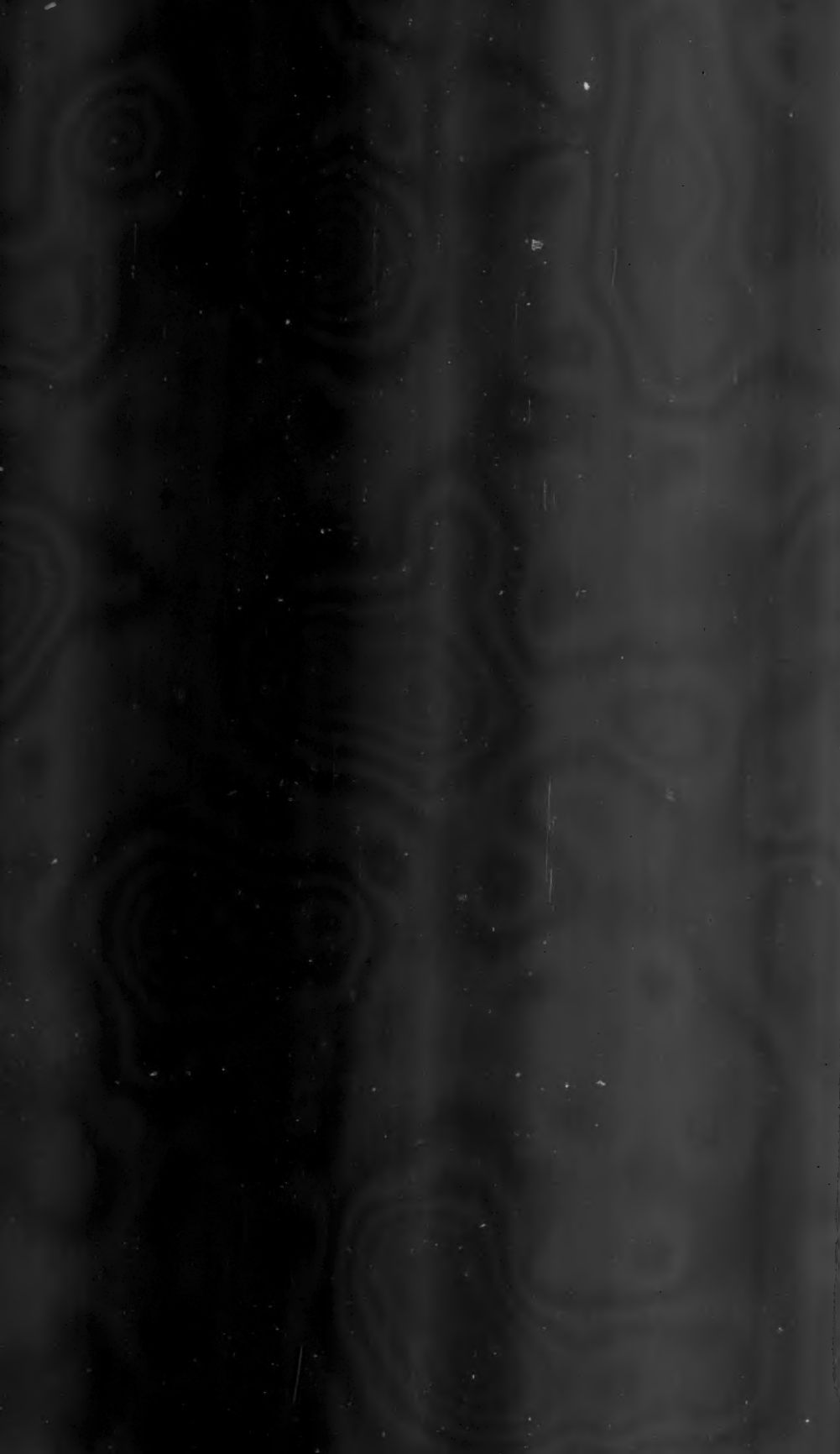
A much closer lithologic resemblance to the Hoback formation is exhibited by the "Paleocene and earliest Eocene" sediments described by Love (1947) in the Jackson Hole area. The two areas at present are separated by the Gros Ventre Range which, however, may have become interposed between the two areas later in the Eocene.

Type locality and stratigraphic sections.—The excellent exposures on the southwestern side of Game Hill, opposite the mouth of Cliff Creek, T38N-R114W, are here designated as the type locality of the Hoback formation (Pl. 4, fig. 1). Section No. 1 of the columnar Sections (Pl. 3) illustrates details of the lithology at the type locality graphically and to scale. This section begins on the southwest side of Game Hill on the northeast side of the Hoback River between Cliff Creek and Shoal Creek. The lowest unit represented is the first well exposed sandstone which crops out, 143 feet vertically above river level. The horizontal course of this section shown between points


Section 1
(A-B)
Section 2
(C-F)
Section 5
(Q-R)
Section 4
(I-J)
Section 3
(G-H)
Section 6
(K-L)
L 1
L 2
L 4
L 7
(X-A)
STRATIGRAPHICAL RELATIONSHIPS OF MEASURED SECTIONS
Scale in feet
2000
1000
500
0



COLUMNAR SECTIONS OF THE HOBACK FORMATION



A and B on the map (Pl. 1) is up the side of Game Hill nearly to its top. The discussion of the lithology of the Hoback formation offered under the heading "General discussion" is applicable to the beds at the type locality. The total thickness of Section No. 1 is 1885 feet. A detailed description of the entire section at the type locality would be too lengthy for inclusion here. Descriptions are given below of units of the section which are representative of the lithologic variants to be encountered throughout the section, or which could be relocated readily in the field. These units can be placed in their proper position in the section on Plate 3 by reference to their numbers. Units 109–113 are included below as representative of a typical cycle of deposition, one which is repeated many times, with minor variation, throughout the section.

Unit No.	Lithologic Description	Thickness (feet)
145	Sandstone, brown; fine-grained; thin, irregular bedding; grading upward into thin-bedded siltstone	2
124	Sandstone, alternate tan and brown layers, weathering tan to brown; conspicuously streaked with iron oxide bands; medium-grained; iron oxide concretions; noncalcareous; deeply wind eroded in pockets; very well exposed	41
113	Sandstone and shale, similar to 111	87
112	Sandstone, medium gray, weathering reddish-brown; fine-grained; arkosic; calcareous	13
111	Sandstone and shale, alternating; poorly exposed on grassy break in slope	85
110	Sandstone, gray to light brown, weathering buff to brown; medium-sized angular grains; arkosic; partly massive, partly thin-bedded; top and bottom calcareous	95
109	Shale, similar to 93	13
93	Shale, dark gray, weathering light gray; nonlaminated; noncalcareous	6
92	Sandstone, light brown, weathering light gray; arkosic; massive; cross-bedded; calcareous; very well exposed	21
82	Sandstone, light brown, weathering to darker brown; medium-sized, angular grains; arkosic; very poorly bedded and massive; calcareous; very well exposed	49
72	Sandstone, medium gray, weathering buff; medium-sized, angular grains; arkosic; poorly bedded; well exposed	41
71	Sandstone, thin, irregularly bedded in ¼–½ inch layers; grades upward into 72 which is more massive; calcareous	10
69	Siltstone, tan, weathering brown; fine-grained; thin-bedded; slightly calcareous; grading upward into sandstone, light gray to light brown, weathering tan; arkosic; massive; slightly calcareous; well exposed	14
57	Sandstone, light gray, weathering buff; medium-sized, angular grains; arkosic; massive; very well exposed	29
55	Sandstone, medium gray; weathering tan; medium-sized, angular grains; thin, crenulated bedding	1
53	Sandstone, dark gray, weathering reddish-brown; massive for most part, some thinly bedded siltstone and shale partings; well exposed	20
51	Limestone, medium gray, weathers tan to brown; iron oxide stained; fine-grained	1
39	Limestone, medium gray, weathering tan; fine-grained; noncrystalline; hackly fracture	4
33	Siltstone, dark gray, weathering light gray; argillaceous; poorly bedded; noncalcareous.	10
32	Sandstone, dark gray, weathering dark reddish-brown; fine-grained; cross-bedded; calcareous	7

Unit No.	Lithologic Description	Thickness (feet)
31	Siltstone, medium brown weathering light gray; non-laminated; noncalcareous	4
23	Sandstone, less arkosic than 22; massive; weathering to spheroidal surfaces; noncalcareous; well exposed	26
21	Shale, dark brown, weathering gray; nonlaminated; noncalcareous; grading upward into siltstone	13
12	Limestone, gray-green, weathering tan-banded	1
8	Shale, dark brown to black, weathering gray; poorly laminated; grading upward into siltstone	11
7	Sandstone, arkosic, and siltstone, light gray, weathering buff; thin, crenulated bedding ½–1 inch thick with some more massive layers 2–3 inches thick	20
5	Sandstone, light gray, weathering buff; medium-sized, angular grains; arkosic; cross-bedded; 8–12 inch layers alternating with thinner bedding; slightly calcareous; weathers spheroidally; well exposed	36
1	Sandstone, light gray, weathering buff; medium, angular grains; arkosic; cross-bedded; well exposed	7

Several other detailed sections were measured. Together with Section No. 1 they form a composite section of the Hoback formation from near the front of the thrust sheet along the western margin of the basin, eastward almost to the uppermost exposed beds.

Section No. 2 begins at point C on the map, about 600 feet vertically above the Hoback River level and one-third mile east of the bridge on U. S. Highway No. 187 crossing Cliff Creek near its junction with the Hoback River. Unit No. 1 of this section was found to be equivalent to Unit No. 139 of Section No. 1 by their continuity in the field.

Section No. 3 was measured along course G-H as shown on the map (Pl. 1). Unit No. 1 of this section is the next unit above Unit No. 78 of Section No. 2. Unit No. 12 of this section lies directly below Unit No. 1 of Section No. 4. (See stratigraphic columns, Pl. 3.)

The highest unit of Section No. 3, i.e. Unit No. 12, lies immediately below Unit No. 1 of Section No. 4. The horizontal position of this section is shown between points I-J on the map (Pl. 1). Fossil locality No. 7, shown on the map, is located 20 feet below the top of Unit No. 3 of this section.

Section No. 5 begins at ditch level along U. S. Highway No. 187 about 50 yards northwest of the junction of the highway and the Dell Creek road. It passes near, but not exactly through, the beds from which the Dell Creek local fauna at Fossil Locality No. 1 was quarried. The fossil vertebrates at Locality No. 1 are found in a thin limestone which immediately overlies Unit No. 14 of this section, but lenses out before crossing the course of this section which passes within 75 feet of the quarry. Unit No. 18 of this section is equivalent to Unit No. 78 of Section No. 2.

Section No. 6 begins at the east end of the old county road bridge across Dell Creek north of the William T. Faris ranch. Unit No. 7 of this section is at a stratigraphic level about 170 feet below the top of Unit No. 11 of Section No. 3.

In all but one instance it was possible to correlate key horizons between measured sections directly by lithologic continuity. Direct correlation between Section 3 (Course G-H) and Section 6 (Course K-L) was not possible in that manner, so in this instance an approximate correlation was obtained with the aid of a surveyor's level. A level sight in the strike direction from the top of Unit No. 7-Section No. 6 intersected a horizon 170 feet below the top of Unit No. 11-Section No. 3. The inaccuracy of this method of correleation is recognized, but is not considered too great for significant relative results (Pls. 1, 3).

Neglecting overlap, the approximate total thickness of the strata measured was:

X-A	2258±feet
Sec. No. 1 (A-B)	1764±feet
Sec. No. 2 (C-F)	5588±feet
Sec. No. 3 (G-H)	2123±feet
Sec. No. 6 (K-L)	2817±feet
Total	14550±feet

Between 1500 and 2500 feet of higher beds in

the Hoback formation lie above the horizon indicated at Point L on the map (Pl. 1) and below the Pass Peak conglomerate. An accurate figure was not sought since the obscurity of the bedding makes accurate dip determination impossible. It is evident, however, that the total thickness of the Hoback formation is over 15,000 feet.

Pass Peak Conglomerate

Distribution.—Above the Hoback formation lies a great thickness of distinctive conglomerates interfingering with sandstone and some shale. These beds occupy the southeastern part of the basin extending from Jack Creek at the southeastern end of the Gros Ventre Range southward and westward to the Hoback Range (Pl. 1). They lap high up on the side of that range, where their relationship to the marginal thrust structures has already been described. They form the Hoback Basin Rim and some small exposures of the sandstone, shale, and conglomerate may be seen there along U. S. Highway No. 187. The southward and southwestward extent of these beds is not yet known. The deposits west of Granite Creek have been mapped as Pass Peak by Eardley *et al.* (1944) and Horberg *et al.* (1949, Pl. II), but Horberg expressed some reasonable doubt as to whether this is correct.

General discussion.—The most conspicuous features of the conglomerate pebbles are their "pressure markings" (Pl. 4, fig. 2). Many of the pebbles have been fractured and recemented. The conglomerates are of the "Pinyon" type (see Love, 1947). The Pinyon conglomerate pebbles are likewise pressure marked and fractured. In the vicinity of Pass Peak, the conglomerates are at least 1500 feet thick forming the summit of Pass Peak and extending down to stream level along Jack Creek. They have been correlated with the Almy conglomerates (Schultz, 1914, p. 77; Nelson and Church, 1943, p. 153–54, 160), with the Pinyon conglomerate of the Jackson Hole and Yellowstone Areas (Blackwelder, 1915, p. 204). The age of the deposits is not yet certain, but they are undoubtedly younger than the Paleocene Pinyon conglomerate, and very probably are younger than the Almy. Both Eardley (1944) and Horberg *et al.* (1949, p. 190) considered them as being middle Eocene in age. No vertebrate fossils have been reported from them. Horberg *et al.* (1949, p. 190) described the beds as follows:

> "At the type locality near Pass Peak the conglomerate is composed of pebbles averaging about 2½ inches in diameter and has a friable sandstone matrix. Over 75 per cent of the pebbles are of foliated quartzite and the remainder consist largely of chert, vein quartz, schist, gneiss, and quartzitic sandstone. Interfingering beds of sandstone become more conspicuous south of the area where the maximum thickness of the formation may reach 5000 feet (Eardley, 1944). The conglomerates west of Granite Creek consist largely of locally derived materials and include fresh water limestones. Metaquartzite cobbles appear to be absent. The age of these deposits is uncertain. They may be Pass Peak, as shown in Plate 2 and Eardley's map (1944), or they may represent marginal facies of the Hoback formation."

Nelson and Church (1943, p. 160) have proposed the Precambrian of the Gros Ventre Range as the source, but the high percentage of dark-gray, brown, black, and blue-black, polished and well rounded quartzite cobbles seems to be evidence against this. The cobbles being carried by recent streams from the Gros Ventres into the Hoback Basin are quite different, predominantly Precambrian light-colored granite, gneiss and schist, and Paleozoic quartzitic sandstones, tough limestones, and dolomites. It therefore seems likely that the pebbles of the Pass Peak conglomerate came from some more remote source.

Fossil Localities in the Hoback Formation

General Occurrence and Preservation of Fossils

The scarcity of good exposures, the steep dip of the beds, and the fluviatile nature of the sediments make it difficult to find fossils, particularly well-preserved specimens, in the Hoback formation. Except for leaves and other plant remains, no fossils were found in the sandstones. Frequently scraps of bone and molluscan shell fragments were found on the shale exposures, but their steep slope and the frequency of landsliding and mudflow down their faces prevents the accumulation of fossils at the surface and destroys those exposed by weathering. The occasional limestone horizons and the shales immediately above them offer the best opportunity for invertebrate collecting. Discovery of fossil vertebrate localities has resulted almost as much from chance as from persistence.

TABLE 1.—FOSSIL LOCALITIES IN THE HOBACK BASIN

Locality No. Sublette Co., Wyoming (Um-Sub-Wy)	Position relative to columnar sections (Pl. 3)	Location	Approximate level relative to top of Unit 42—Sec. 2 Conglomerate (feet)		Remarks
19		NW¼, Sec. 26, T37N-R113W	Above	+9200	Bone fragments of large mammal
17		S. side Clark Butte, Sec. 23, T37N-R113W	Above	+9200	Bone fragments
5		NE¼, Sec. 35, T38N-R113W	Above	+9200	Hackberry seeds (*Celtis*)
4	Unit 14—Sec. 6	SW¼, Sec. 35, T38N-R113W		+8945	*Lepidosteus*, turtle, alligator, *Esthonyx?* Hackberry seeds
3	Unit 13—Sec. 6	Center E.½, Sec. 27, T38N-R113W		+8690	Hackberry seeds
2	Unit 13—Sec. 6	Center S.½, Sec. 27, T38N-R113W		+8690	*Hyracotherium*, etc. See faunule list
16	Unit 9—Sec. 6	SE¼, NW¼, Sec. 34, T38N-R113W		+7930	*Coryphodon* sp., molar fragment; other bone
13	Unit 9—Sec. 6	NW¼, SW¼, Sec. 27 to NE¼, SE¼, Sec. 28, T38N-R113W		+7800	Fossil wood
7	Unit 3—Sec. 4	NE¼, Sec. 16, T38N-R113W		+7100	*Coryphodon*, sp., *Probathyopsis successor*, turtle
8	Unit 1—Sec. 4	SE¼, Sec. 16, T38N-R113W		+6785	Bone fragments
11		NE¼, Sec. 5, T37N-R113W	Between and	+6380 +7480	Turtle fragments
9	Unit 3—Sec. 6	SE¼, SW¼, Sec. 21, T38N-R113W		+6380	Bone fragments, Mollusc fragments
10	Unit 3—Sec. 6	NW¼, SE¼, Sec. 21, T38N-R113W		+6380	Mollusc fragments
14	Unit 7—Sec. 3	SE¼, Sec. 17-NE¼, Sec. 20, T38N-R113W		+4900	Fossil wood
1	Unit 15—Sec. 5	NE¼, NW¼, Sec. 29, T38N-R113W		+4030	Vertebrates, see faunal list
12		SW¼, Sec. 31-T38N, R113W to SW¼, Sec. 6, T37N-R113W	Between and	+3500 +4264	Bone fragments, Molluscs (see Schultz, 1914, p. 70)

TABLE 1.—FOSSIL LOCALITIES IN THE HOBACK BASIN—*Continued*

Locality No. Sublette Co., Wyoming (Um-Sub-Wy)	Position relative to columnar sections (Pl. 3)	Location	Approximate level relative to top of Unit 42—Sec. 2 Conglomerate (feet)		Remarks
15	Unit 42—Sec. 2	Road level near V-V Ranch U. S. 187		0	Fossil wood
18	Unit 13—Sec. 2	N. of Hoback R. near river level ¼way bet. Cliff Cr. and V-V Ranch		−1393	Fossil leaves
6		300 yds NE of jct Cliff & Kerr Cr., T38N-R114W	Between and	−4750 −5069	Molluscs, *Lepidosteus* indet. fish

Locality List and Locations

Table 1 is a complete list of the localities within the Hoback Basin at which fossils, regardless of their completeness or state of preservation, were found during the course of this study. The localities are arranged stratigraphically. Their approximate level is indicated above (+) or below (−) the conspicuous conglomerate (Unit 42 of Section No. 2. See Pl. 3) exposed at road level along U. S. Highway No. 187 near the V-V Ranch. These levels are estimates and are only relatively correct. The conglomerate horizon was selected as a datum since it can easily be relocated for future use as a reference horizon and since the lowest beds exposed in front of the marginal thrust sheet are not everywhere at the same horizon. All the localities listed occur within Sublette County and are referred to in the table and elsewhere as UM-Sub-Wy localities. Localities 5, 17, and 19 lie in higher horizons than were reached by the measured sections.

FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

General Statement

The author has collected remains of fossil vertebrates from the Hoback formation at ten localities (Table 1) in Sublette County, Wyoming. Only six of these, however, University of Michigan-Sublette County-Wyoming (UM-Sub-Wy) Localities 1, 2, 4, 6, 7, and 16, have furnished identifiable specimens. The faunules are discussed in ascending stratigraphic order from the lowermost to uppermost. For the location and stratigraphic position of the beds at these localities see map and Table 1. Unless otherwise stated, all specimen numbers are those of the University of Michigan Museum of Paleontology (U.M.M.P.).

UM-Sub-Wy Locality No. 6

The beds at Locality No. 6 occur in the lowest horizon in the Hoback formation from which fossils were collected (Table 1). The writer was directed to this locality by Dr. A. J. Eardley. A large, but as yet undescribed, collection of freshwater molluscs from this locality has been assembled. The fossils occur in several impure, argillaceous and petroliferous limestone beds which are exposed in the side of an old logging dugway about 300 yards northwest of the junction of Cliff and Kerr creeks. Minute charaphyte oögonia are dispersed in great quantities throughout the rock and many of the cavities originally occupied by these plant remains are now filled with an asphaltic residue. The only vertebrate remains found here were a gar pike scale, U.M.M.P. No. 27507 (*Lepidosteus*) and a shattered fragment of an undetermined fish dentary (U.M.M.P. No. 27508). The age of the beds at this horizon is not known, but they lie more than 8000 feet below the Dell Creek Quarry, UM-Sub-Wy locality No. 1, from which a Tiffanian mammalian faunule was collected.

UM-Sub-Wy Locality No. 1, The Dell Creek Quarry

Occurrence.—Fossil vertebrates were first discovered at this locality by the author in 1947. Invertebrates had been collected there prior to that by faculty and students from the University of Michigan Camp Davis. The majority of the vertebrate specimens have been recovered from an 8-inch layer of impure silty limestone of limited extent which occurs in the top of Unit 15—Section 5 (Pl. 3) at the quarry (Pl. 4, fig. 3). Some of the material comes from the lower 3 to 4 inches of shale immedi-

ately above the limestone. Large numbers of gastropods, including members of the genera *Physa*, *Planorbis*, and *Oreohelix*, occur in close association with the vertebrate material in this quarry. Minute fragments of bone occur throughout the limestone, but no complete postcranial skeletal elements have been found. The bone fragments usually appear freshly broken. The majority of the vertebrate specimens consist of isolated mammal teeth and skull and jaw fragments. The fragments do not appear waterworn and the teeth are usually well preserved. The teeth and bones are dark brown to black.

Class REPTILIA

Order CROCODILIA

Family ALLIGATORIDAE

Allognathosuchus sp.

Seven isolated teeth and a fragment of one of the dermal roofing bones of the skull, all found rather closely associated, U.M.M.P. No. 27236, are referred to this rather common early Cenozoic crocodile. The blunt, vertically ribbed teeth are quite characteristic of this genus.

Order SQUAMATA

Suborder LACERTILIA

Family indet.

Haplodontosaurus excedens? (Gilmore, 1928)

(Pl. 5, figs. 1, 2)

Harpagosaurus excedens GILMORE, 1928, Nat. Acad. Sci., Mem., vol. 22, no. 3, p. 157, fig. 99.
Haplodontosaurus excedens (GILMORE) GILMORE, 1942, Am. Philos. Soc., Pr., vol. 85, no. 2, p. 164–65, figs. 10; 12.

A fragment of a right dentary, U.M.M.P. No. 27217, as indicated by the convex outer surface of the bone and the distribution of the mental foramina, is referred to this genus and questionably to this species. This specimen was collected in the summer of 1948 by the University of Michigan field party. Fragment preserves parts of 9 rather acutely pointed, fully pleurodont teeth with spaces for 4 others bringing total to 13. Undoubtedly several teeth anterior to these were present in the missing portion of the dentary. Teeth appear to be slightly flattened fore and aft as in the type of *H. excedens*. Comparison with Princeton Museum specimen No. 14567, part of a left dentary from Silver Coulee beds of Wyoming identified by Gilmore (1942, p. 165) as pertaining to *H. excedens*, was so close as to provide no visible grounds for separation either on basis of size or tooth characters. Specimen, however, is too fragmentary for more than doubtful specific reference. Type of *H. excedens* was described from Lebo beds of Montana. The teeth in No. 27217 are more acutely pointed than in *Harpagosaurus silberlingi* Gilmore from the Lebo formation of Montana and less so than in *H. parvus* Gilmore from the Lance, two otherwise closely similar forms.

Lacertilian?, indet.

A specimen, collected at the Dell Creek Quarry by Mr. H. P. Zuidema in company with the author in 1947, may be a portion of a lizard jaw, but the specimen is too poorly preserved for identification.

Class MAMMALIA

Order MULTITUBERCULATA

Family PTILODONTIDAE

Genus ***Ptilodus*** Cope, 1881

Ptilodus COPE, 1881, Am. Natural., vol. 15, p. 921–22

Ptilodus? fractus n. sp.

(Figs. 2, 3; Pl. 5, figs. 3, 4)

HOLOTYPE: No. 27219, University of Michigan Museum of Paleontology, a left P_4–M_1, collected by the University of Michigan 1948 field party.

PARATYPES: No. 27222, a left P_4 with posterior part missing; No. 27225, a left P_4; No. 27226, a right P_4 with posterior part missing; No. 27227, a right P_4, with anterior edge including at least first serration missing; No. 27228, a left P_4 with posterior part missing.

HORIZON AND TYPE LOCALITY: Upper Paleocene, Dell Creek local fauna, Hoback formation. UM-Sub-Wy Locality No. 1, Dell Creek Quarry, Dell Creek road near U. S. Highway No. 187, Sublette County, Wyoming.

DIAGNOSIS: A ptilodontine with P_4 smaller than in *Ptilodus mediaevus* Cope, *P. montanus* Douglass, or *P. wyomingensis* Jepsen. Lateral contour of P_4 closest to *Ptilodus*, but differs from these 3 species in that the contour of the anterior portion of the crest of the tooth is not smoothly convex but presents a sharp increase in the rate of backward curvature between the anterior basal concavity and the first serration about a third of the distance below the latter. Disposition and number of ridges descending from the first serration as on P_4 in *Ptilodus*, but with middle ridge terminating at, or dividing around, a shallow, narrow, more or less well-developed groove. This groove begins at the "break" in the anterior contour of the crest and broadens only slightly as it descends to the basal concavity. The specific name, *fractus*, is given to this species in reference to the "break" in the anterior lateral contour of P_4.

DESCRIPTION OF HOLOTYPE: Specimen No. 27219 consists of a left P_4–M_1. The two teeth are not held together in a jaw fragment, but were found closely associated in the matrix in relative positions almost

exactly the same as they would have occupied in the jaw. There is no doubt that they belonged to the same individual. Convex lateral contour (outline of tooth as viewed from lingual side) of P_4 is low, rounded, and very nearly symmetrical, posterior portion of crest being only slightly higher than between "crest" and "middle ridge" should be kept in mind when using Jepsen's (1940, p. 264–65) "Synopsis of Generic Characteristics of American Tertiary Ptilodontidae." Lower two-thirds of anterior face of P_4 below first serration is not a thin blade, but occupied by a narrow, shallow groove,

FIGURE 2.—*Ptilodus? fractus* N. SP.
Superposed outlines of P_4 of holotype and four paratypes.

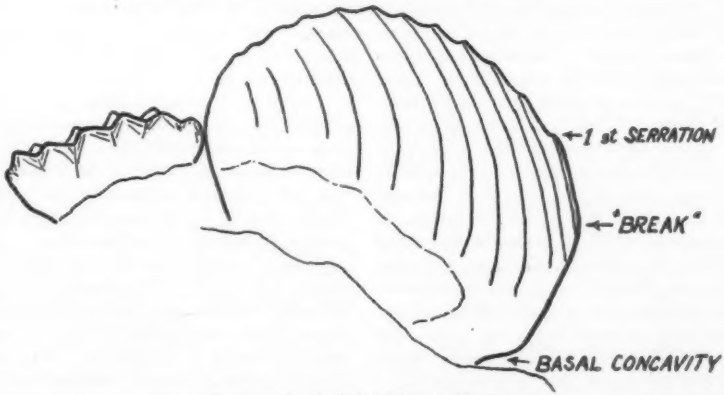


FIGURE 3.—*Ptilodus? fractus* N. SP.
Holotype, U.M.M.P. No. 27219. Lingual view, left P_4–M_1, ×11. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming.

anterior portion. A distinct anterior basal concavity is present. Symmetry of lateral contour is imperfect due to character of anterior portion of the crest between anterior basal concavity and first serration. Crest of tooth rises steeply above anterior basal concavity for about two-thirds of the distance toward first serration. At this point there is an abrupt "break" in the slope. Thereafter crest rises less steeply, with an increased backward component, to first serration.

There are three equally spaced ridges from first serration on P_4. The middle one follows, or perhaps is identical with, thin crest of tooth as it descends antero-ventrally from first serration. This identity appearing at the level of the break in anterior lateral contour and descending to anterior basal concavity, widening only slightly in this distance.

The two lateral ridges from first serration descend only about a third of the distance toward basal concavity, terminating, without branching, at the level of the break in antero-lateral contour. The middle ridge, on this specimen, can be considered either as terminating at apex of anterior groove or as dividing into two less distinct, symmetrically spaced ridges which continue downward as lips of groove.

There are 12 serrations on P_4. The ridges descending from these serrations on lingual and labial faces of tooth are unbranched. Distance from center of

anterior basal concavity to first serration is equal to a half the greatest length of P_4 (see Jepsen, 1940, p. 265–66).

The presence of an anterior basal concavity on P_4 indicates that a P_3 was probably present. The probability of this is heightened by what appears to be the outline of a tooth root in the matrix immediately in front of P_4. If this is the root of P_3, that tooth was relatively large in comparison with P_4.

M_1 is distinctly *Ptilodus*-like with a few large cusps. Cusp count is 6:5 (6 cusps on the labial row and 5 on the lingual row), 2 most posterior cusps on lingual row merging for about half their height. One posterior outer cusp was lost in preparation so that a count of the remaining cusps would be misleading. P_4 is relatively shorter than in other species of *Ptilodus*, being less than twice the length of M_1.

Description of paratypes.—Three left and two right isolated lower fourth premolars from same level and locality as holotype, and all found within 25 feet, are designated as paratypes (Table 2). Figure 2 shows outlines of P_4 of the holotype and paratypes; To facilitate comparison with the lateral contours of other ptilodontids as figured by Jepsen (1940, p. 250–66), these were drawn to scale so that the greatest length of P_4 is the same as in that paper.

No. 27222, anterior portion only of a left P_4 back to fourth serration, differs from holotype in that anterior groove below first serration is less pronounced and lies off exact midline of tooth somewhat toward lingual side. Middle ridge from first serration, in contrast to that of the type, continues strongly down labial side of groove for about half the length of the groove. No indication of a ridge can be seen along lingual edge of groove, and lateral ridge from first serration on lingual side is deflected posteriorly near its lower end so as to join the corresponding ridge from second serration. Otherwise this specimen agrees with the holotype.

No. 27225, a complete left P_4 showing wear on posterior part of crest differs from the type only in possessing a shallower, less distinct anterior groove, and in that the lateral ridge from the first serration, on the lingual side, joins the ridge from the second serration as in No. 27222.

No. 27226, a right P_4 lacking only a small portion posteriorly, compares with holotype in same manner as No. 27222.

No. 27227, a right P_4 lacking first serration and anterior face is inseparable from holotype in size and in characters of remaining portion.

No. 27228, anterior half of a left P_4 is indistinguishable from holotype in its remaining portion.

REFERRED SPECIMENS: A left M^1, No. 27220; fragments of 4 ptilodontine incisors, No. 27509; 4 isolated upper second or third premolars, No. 27510; and a transversely distorted fragment of a palate with left P^{1-3}, part of right P^1 and the roots of right P^3, No. 27221, are referred to this genus and species with some question, on the bases of size and provenience. Although the referred specimens are not directly comparable with the holotype, all are from the same locality as the holotype.

Some of the posterior cusps are missing from the outer row on the specimen of M^1, but the cusp count appears to have been 9?:10:6 (Cusp rows counted from labial to lingual side). Dimensions of this tooth are: length, 4.9 mm.; width, 2.2 mm.

The 4 isolated upper premolars are all closely similar to one another and to the left P^2 of specimen No. 27221. All are 4-cusped. See Table 2 for dimensions and cusp counts on No. 27221.

Although not complete, the 4 incisors are ptilodontine in nature. They are slender, only slightly compressed laterally, roughly elliptical in cross section, and with a longitudinal groove on the internal face. The crowns are completely covered with enamel. The fact that no other type of ptilodontid incisor is included in the collection is indirect evidence that neither of the 2 ptilodontid genera represented by other teeth in the collection are eucosmodontines.

DISCUSSION: Ptilodontid lower fourth premolars are large, very distinctive, and by far the commonest ptilodontid teeth in collections. The value of this tooth in taxonomic studies of the ptilodontids has been fully discussed by Jepsen (1940, p. 249–50). There seems to be no reason to doubt that the characters exhibited by the holotype and paratypes are those of a distinct species. The close similarity of the lateral contour of P_4, a character already used with much success in multituberculate taxonomy by Jepsen (1940), demonstrates the unity of the group when considered in conjunction with the other distinctive structural details described above. There are both right and left lower fourth premolars in the collection, but none can be proven to belong to the same individual, nor is their sex determinable. If they are assumed to belong to 6 different individuals, the superposition of their lateral outlines (Fig. 2) gives some measure of individual variation in the shape of P_4 within the species. The description of paratypes shows them to differ only slightly in characters of the ridges from the first serration and in the groove on the front edge of P_4.

Direct comparisons were made with the types of all species belonging to genera which could not be immediately eliminated from consideration by use of Jepsen's (1940) synopsis of the characteristics of the ptilodontids or by reference to published descriptions of genera.

Aside from the fact that none of the ptilodontid

incisors in the collection is of the eucosmodontine type, the species bears only slight resemblance in what are considered generic characters to any of the individual genera of that subfamily. P_4 is longer than M_1 as in *Eucosmodon*, but also as in several greater distance from the anterior basal concavity to the first serration relative to the greatest length of P_4. The specimens are too few and incomplete to warrant any speculations as to the phylogenetic significance of this new species.

TABLE 2.—DIMENSIONS (IN MM.) AND RATIOS OF TEETH OF *Ptilodus? fractus* N. SP.

U.M.M.P. Specimens	P_4				M_1				$\frac{LP_4}{LM_1}$
	Serrations	L	W	$\frac{L}{W}$	Cusps	L	W	$\frac{L}{W}$	
Holotype									
27219	12	5.6			6:5	3.2	1.5	2.13	1.75
Paratypes									
27222			2.8						
27225	12	5.7	2.4	2.38					
27226	10+		2.8						
27227	11+		2.5						
27228	9+		2.4						
Means	12	5.65	2.58	2.19					1.76

Referred Specimens	P^1			P^2			M^1		
	Cusps	L	W	Cusps	L	W	Cusps	L	W
27220							9?:10:6	4.9	2.2
27221	3	2.3	1.8	4	2.4	2.0			

ptilodontine genera. There is no close resemblance to *Pentacosmodon*, and the similarity to *Neoliotomus* and *Microcosmodon* lies only in the small number of cusps on M_1. The group is therefore considered as more properly placed in the subfamily Ptilodontinae Jepsen (1940, p. 262–63).

The structural resemblances and differences between the 6 American Paleocene ptilodontine genera from which lower teeth are known and *Ptilodus? fractus* were considered in detail. The closest resemblance was found to be to the genera *Ptilodus* Cope, 1881, and *Prochetodon* Jepsen, 1940 (Table 3); it will be noted that some of the resemblances are common to both genera. The species is tentatively assigned to the genus *Ptilodus* on the grounds of the close similarity in lateral contour, number and distribution of ridges from the first serration, and number of serrations on P_4, and also because of the small number of cusps on M_1. It is distinguished from *P. mediaevus*, *P. montanus*, and *P. wyomingensis* by the "break" in the convex anterior slope and the groove on the front of P_4, by its smaller size, and by the smaller ratio of the length of P_4 to length of M_1. It is also distinguished by the

Ectypodus powelli Jepsen

(Pl. 5, fig. 5)

Ectypodus powelli JEPSEN, 1940. Amer. Philos. Soc., Pr., vol. 83, no. 2, p. 308–09, fig. 18, pl. III; figs. 1, 1a.

A single left P_4 of a small ptilodontine, U.M.M.P. No. 27244, is referred to this genus and species. Direct comparison with the holotype revealed no distinguishable differences. Lateral contour of tooth is low, and higher in front than in rear. There are 10 serrations, with 3 asymmetrically spaced ridges diverging from first serration. Length of specimen is 3.4 mm., width 1.4 mm.

Four other isolated teeth as listed may questionably pertain to this form also: anterior part of a small ptilodontine lower incisor, No. 27511, and 3 small upper first molars, Nos. 27218, 27223, 27229, too badly broken to obtain reliable cusp counts. Incisor is slender, ovaloid in cross section, with an enamel covered crown. It is much smaller than incisors referred to *Ptilodus* from the same quarry. The 4 upper molars are larger than those of *Ectypodus laytoni* Jepsen. No. 27218 measures 2.9 mm. long and 1.3 mm. wide. This is close to the dimen-

sions (2.8 mm. and 1.4 mm.) of an M^1 (Princeton No. 14463) in the Princeton collection from the Silver Coulee beds, which Jepsen (1940, p. 309) referred with some doubt to *E. powelli*.

crown. There is a talonid cuspule posterior to main cusp. No anterior cuspule is present. P_3 is broken. Only anterior and posterior cuspules, which are distinct and part of main cusp, are preserved. The

TABLE 3.—RESEMBLANCES AND DIFFERENCES OF *P? fractus* N. SP. AS COMPARED WITH THE GENOTYPES OF *Ptilodus* AND *Prochetodon*

	Resemblances	Differences
Ptilodus	1. P_4 relatively low and rounded. 2. Lateral contour of P_4 comparatively symmetrical. 3. Symmetrical distribution of ridges from 1st serration on P_4. 4. Height of crown of P_4 compared to length. 5. M_1 with few cusps. 6. Number of serrations on P_4.	1. Groove on front of P_4. 2. Anterior edge of P_4 not smoothly convex, but with a "break". 3. Smaller than other species of *Ptilodus*. 4. P_4 less than twice the length of M_1. 5. Distance from anterior basal concavity to 1st serration equal to ½ length of P_4.
Prochetodon	1. P_4 relatively low. 2. Sub-Convex above anterior basal concavity. 3. Symmetrical distribution of ridges from 1st serration on P_4. Short middle ridge in holotype. 4. Groove on front edge of P_4. 5. Number of serrations on P_4.	1. Lateral contour of P_4 symmetrical. Front not conspicuously lower than rear. 2. Front edge of P_4 rises more steeply; contour "breaks." 3. P_4 shorter in relation to M_1. 4. Groove on front of P_4 narrower, shallower, more medial in position; not rising as high. No median ridge in center of groove. 5. Anterior basal concavity not as broad. 6. Four cusps on P^3?.

Order INSECTIVORA

Family LEPTICTIDAE

Leptacodon packi Jepsen, 1930

(Pl. 5, figs. 6, 7)

Leptacodon packi JEPSEN, 1930. Am. Philos. Soc., Pr., vol. 69, no. 7, p. 510–511, pl. 8; figs. 4–5.

An incomplete right ramus, No. 27235, pertains to this genus and probably to this species. Lower dental formula of this specimen was 3-1-4-3. Only P_2 is complete. P_3 and P_4 are broken. The presence of the remaining teeth is indicated by alveoli.

There are 3 incisive alveoli. I_{1-2} appear to have been laterally compressed, I_3 more nearly circular. The incisors were somewhat transversely placed in relation to the other teeth in the jaw. The alveoli of the large canine and slightly smaller P_1 are antero-posteriorly elongate ovals. P_1 was single rooted. P_2 is double rooted with a short, high, trenchant cavity in the matrix which was occupied by the main cusp is sufficient to show that this tooth was also trenchant, the same height but longer than P_2, shorter and lower than P_4. P_4 was submolariform with a separate, relatively high paraconid and a distinct protoconid and metaconid. Tip of protoconid was irreplaceably broken during preparation, but this cusp was larger, somewhat higher, and antero-external to metaconid. Talonid is small and bears 3 cusps. Of these, hypoconulid is medially placed and nearly confluent with hypoconid. Rounded alveoli of M_{1-2} are similar to one another in size and shape. Posterior root of M_2 is triangular in outline with apex toward rear.

Coronoid process rises abruptly and steeply. There are 2 mental foramina, a higher one beneath anterior root of P_2 and a slightly lower one between anterior root of P_4 and posterior root of P_2.

Specimen is smaller than *Leptacodon ladae* Simpson and close to *L. tener* Matthew and Granger and

L. packi in size. It was found to be inseparable from the holotype (Princeton No. 13296) of *L. packi* in character of basined talonid and height of protoconid on P_4, and in position of mental foramina. Dimensions of specimen are: I_3 to M_3 (alveolar length), 11.5 mm.: M_1 to M_3 (alveolar length), 4.7 mm.

Order PRIMATES

Family PLESIADAPIDAE

Genus *Plesiadapis* Gervais, 1877

Plesiadapis GERVAIS, 1877. Jour. Zool., vol. 6, p. 74–79.
Nothodectes MATTHEW, 1915. Am. Mus. Nat. Hist., Bull. vol. 34, Art. 14, Part 4, p. 482–83, fig. 52.

Plesiadapis farisi n. sp.

(Fig. 4; Pl. 6, figs. 1, 2)

HOLOTYPE: No. 27230, University of Michigan Museum of Paleontology. A right maxillary with P^3–M^3, collected by George R. L. Gaughran of the University of Michigan field party, summer of 1948.

HORIZON AND TYPE LOCALITY: Upper Paleocene, Dell Creek local fauna, Hoback formation. UM-Sub-Wy Locality No. 1, Dell Creek Quarry, Dell Creek road near U. S. Highway No. 187, Sublette County, Wyoming.

DIAGNOSIS: A plesiadapid near *Plesiadapis gidleyi* (Matthew) in size. Paracone and metacone confluent on P^{3-4}. Protoconule on these teeth is a large, distinctly separate, anteroposteriorly elongate cusp. P^{3-4} and M^2 are significantly wide in relation to M^1. Transverse P^3 is exceptionally wide in relation to its length. P^4 is almost as wide as M^1. The specific name, *farisi*, is given to this new species in grateful acknowledgment of the many kindnesses, generous advice, and assistance rendered to the writer and to the University of Michigan field parties by Mr. and Mrs. William T. Faris of Bondurant, Wyoming, and members of their family.

DESCRIPTION OF HOLOTYPE: Dentition on holotype is in an early stage of wear. Paracone and metacone are confluent on P^{3-4}. Only a slight groove limited to upper part of ridge formed by these cusps suggests that both are present. The protoconules on these teeth are robust, anteroposteriorly elongate, and distinctly separate from paracone. There is a distinct parastyle on P^3 and there apparently was also one on P^4 although it has been broken. A very weak metastyle is present. The labial surfaces of premolars are rugose. Posterior cingulum was evidently complete on premolars, but part of it has chipped from P^4. External cingulum is incomplete on premolars. Anterior cingulum is complete on P^4, but incomplete on P^3. Both premolars are transverse. Width of P^3 in relation to length is especially great for this tooth. P^4 is very nearly as wide as M^1. A crest joining protocone to parastyle passes anterior to protoconule on premolars.

Paracone, metacone, and protocone are all distinct molar cusps. M^{1-2} have well-developed protoconules, but metaconule is less distinct. Presence of a hypocone on the molars is very doubtful and posterior cingular shelf does not extend internally beyond protocone. External cingulum is large and complete on molars and broadens inwardly to form a short ridge between para- and metacones in mesostylar area. It is particularly stout on anteroexternal corner of M^2. Anterior cingulum extends lingually on M^1 to a point opposite protocone, but does not extend upward to join this cusp. On M^2, anterior cingulum does not extend quite as far inwardly. On M^{1-2} anterior and posterior wings from the protocone connect this cusp with the conules. M^1 is relatively narrow and long. It is reduced with respect to premolars and M^2, and is more quadrangular. The transverse M^2 is unusually wide in relation to its length, broadly and deeply emarginate labially, and shows a strong anterior deflection of its internal base. Dimensions and ratios of the holotype are given in Table 4 and compared there with 2 other species of the genus *Plesiadapis*.

The infraorbital foramen is situated above P^3 (5.5 mm.), a plesiadapid characteristic cited by Jepsen (1934, p. 300). The zygomatic arch arises above M^1. Between anterior root of zygomatic arch and infraorbital foramen, the maxillary is deeply and broadly excavated.

REFERRED SPECIMENS: A left P^3, No. 27512, and a right P^3, No. 27513, are very close to that of the holotype. Paracone and metacone are indistinguishable on the left premolar and separated only slightly on the right premolar. They are identical with the holotype in other respects.

A broken upper incisor, No. 27514, possibly pertains to this form as does a fragment of an M^2, No. 27515. A P_4, No. 27516, may likewise belong to this species.

No. 27517, a right M_3, occludes perfectly with the upper dentition of the type and, if its reference to *P. farisi* here is correct, it offers a further means of distinction between this species and *P. gidleyi*. Comparing (Table 5) the dimensions of this tooth with the mean of a series of the same tooth in specimens of *P. gidleyi*, as analysed by Simpson (1935, p. 6) shows that No. 27517 belonged to a plesiadapid almost certainly distinct from *P. gidleyi*. Further comparison to Simpson's data (1935, p. 7) indicates that the form's distinctness from *P. gidleyi* is of a different character and degree than is that of the other species analysed by Simpson.

A broken left M^3, No. 27518, is also referred to this genus and species. It is very slightly larger than that tooth in the holotype. The inner base is more asymmetrical, and the posterior cingular shelf somewhat more expanded.

FIGURE 4.—*Plesiadapis farisi* N. SP
Holotype, U.M.M.P. No. 27230. Portion of maxillary with P^3–M^3, labial and occlusal views, ×6.

DISCUSSION: *Plesiadapis farisi* is a distinctly smaller form than the European species. It is also smaller than *P. cookei* Jepsen, from the Clark Fork beds of Wyoming. In addition, *P. cookei* lacks the conule of P^4. Another Clark Fork species, *P. dubius* (Matthew), is known only from the lower dentition. Direct morphological comparison being impossible, an attempt to occlude the lower dentition of the holotype of *P. dubius* with the upper dentition of the holotype of *P. farisi* proved the latter species to be somewhat larger and probably distinct. Simpson (1937b, p. 168–69) has suggested that the two species from the Melville formation of Montana, *P. rex* Gidley and *P. anceps* Simpson, may be synonymous. The former is known only from the holotype, an isolated M_2. *P. anceps* lacks the conule on P^3, has it only weakly developed on P^4, and lacks mesostyles on the upper molars, all features which, in distinction, are well developed on *P. farisi*. The holotype of *P. rubeyi* Gazin, a partial lower dentition from the Almy formation in Wyoming, is that of a larger form. A broken M^3, U.S.N.M. No. 16697, in the collections of the U. S. National Museum, is from the same area and is referred on the label to *P. rubeyi*. It is probably larger than the M^3 in *P. farisi* would have been if it were preserved in the holotype.

P. farisi compares most closely with *P. gidleyi* from the Tiffany and *P. fodinatus* Jepsen from the Silver Coulee local faunas.

As compared with *P. fodinatus*, paracone and

TABLE 4.—DIMENSIONS (IN MM.) AND RATIOS OF TEETH OF THE HOLOTYPE OF *Plesiadapis farisi* N. SP. Compared with those of *P. gidleyi* and *P. fodinatus* Jepsen

Specimens	P^3			P^4			M^1			M^2		
	L	W	$\frac{L}{W}$	L	W	$\frac{L}{W}$	L	W	$\frac{L}{W}$	L	W	$\frac{L}{W}$
Plesiadapis farisi U.M.M.P. 27230 type	2.6	3.5	.743	2.7	4.3	.628	3.2	4.4	.727	3.5	5.4	.648
Difference between *P. farisi* and mean of *P. gidleyi*			.144			.032			.011			.004
Plesiadapis gidleyi Means of A.M.N.H. Nos. 17170 (type), 17173, 17200	2.35	2.65	.887	2.33	3.53	.660	3.1	4.2	.738	3.13	4.8	.652
Difference between *P. fodinatus* and *P. farisi*			.019			.047			.049			.089
Plesiadapis fodinatus P.U. 13306 Allotype of Jepsen	2.9	3.2	.906	2.9	4.1	.707	3.7	4.7	.787	4.0	5.4	.741
Difference between *P. fodinatus and P. farisi*			.243			.079			.060			.093

TABLE 5.—DIMENSIONS (IN MM.) OF RIGHT M_3 OF *Plesiadapis farisi*? N. SP. (REFERRED SPECIMEN) Compared with M_3 of *P. gidleyi*

Specimen	U.M.M.P. No. 27517 *P. farisi*? Ref. spec.	*P. gidleyi* (means)	Difference	Standard Deviation of *P. gidleyi*	Difference Divided by Standard Deviation	Probable degree of Distinctness
Greatest length	5.2	4.63	.57	.12	4.75	Practically certain
Greatest width	3.03	2.87	.16	.13	1.23	Not probable

metacone are less distinct on the premolars in *P. farisi*. Protoconule is stronger, less elongate, and more distinct. The transverse P^3 is wider in relation to its length and more nearly rectangular in outline with the posterior margin of the lingual half of the tooth less deeply indented. P^{3-4} and M^2 are distinctly wider in relation to M^1. Jepsen (1930, p. 516) described P^3 on the allotype of *P. fodinatus* (Princeton No. 13306) as lacking the protoconule.

Compared with *P. gidleyi*, P^3 in *P. farisi* is again wider in relation to its length. P^{3-4} and M^2 are wider in relation to M^1. The protoconule on the premolars is stouter, more elongate, and more distinctly separate from the paracone. The paracone and metacone are more confluent. The external cingulum on M^{1-2} is stronger. The inner molar base is more nearly square on M^1 and less nearly square on M^2 than in *P. gidleyi*.

No definite statement as to the stratigraphic position of *P. farisi* is felt to be warranted by the material here described. It is probably more closely related to the Tiffanian than to the Clarkforkian plesiadapids. Simpson's observation (1935, p. 27) that the conule on P^{3-4} is often reduced or absent in the more advanced species of *Plesiadapis* may be of some significance here.

Plesiadapis sp.

(Pl. 5, fig. 8)

An isolated right P^4, No. 27519, is referred to this genus, but belonged to a species distinct from *P. farisi* from the same horizon and locality. The inner base of the tooth is missing so it is impossible to determine the width of the tooth or to observe the nature of the protocone. The tooth is distinctly longer anteroposteriorly than P^4 in *P. farisi*. The paracone and metacone are broadly separated and equal in size and height. Their separation is continued downward on the labial side of the tooth as a groove reaching the external cingulum. The external cingulum is strong and complete. There is a well-developed parastyle. The protoconule is larger, more distinctly separate from the paracone, and even more anteroposteriorly elongate than in *P. farisi*. Although this tooth suggests the presence of two species of the same genus in this quarry fauna the two forms are apparently so distinct that they afford no reason for questioning the possible validity of Cabrera's "Law of Ecologic Incompatability" (Simpson, 1937b, p. 64–67).

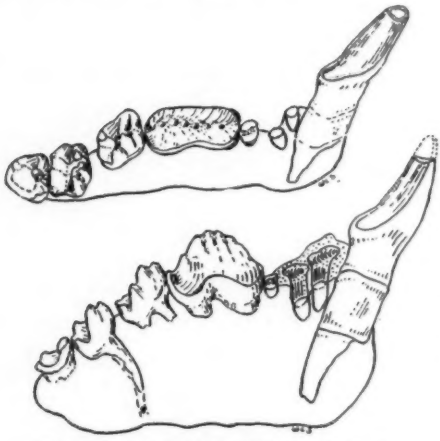


FIGURE 5.—*Carpodaptes hobackensis* N. SP.
Holotype, U.M.M.P. No. 27233. Occlusal and lingual views of left ramus with I—M_3, ×6. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming.

Plesiadapid, gen. and sp. indet.

(Pl. 5, fig. 9)

A small primate left upper incisor, No. 27520, with 3 apical cusps and a strong posterobasal cusp probably is that of a plesiadapid. It is, however, too small to have belonged to either of the 2 forms described above, comparing more closely with *Pronothodectes* Gidley.

Family CARPOLESTIDAE

Genus ***Carpodaptes*** Matthew and Granger, 1921

TYPES: *Carpodaptes aulacodon* MATTHEW AND GRANGER, 1921, Am. Mus. Novit., no. 13, p. 6.

DISTRIBUTION: Tiffanian of Colorado, Wyoming and Montana.

DIAGNOSIS: (Modified after Simpson, 1935, p. 10) Dentition $\frac{1}{1}$-1-3-$\frac{3}{3}$. Canine? and P_{3-4} much reduced with single, vertical, cylindrical roots and small knob-like crowns. P_4 more enlarged than in *Elphidotarsius* Gidley, but less so than in *Carpolestes* Simpson, with 5 to 6 more or less well-differentiated serrations on main blade, and a single talonid cusp distinct from main blade and lower than trigonid of M_1. Trigonid blade of M_1 slightly less elongate than in *Carpolestes*. Paraconid and metaconid are equidistant anteroposteriorly from protoconid. Third lobe of M_3 relatively smaller, less projecting, and less asymmetrical than in *Carpolestes*.

Carpodaptes hobackensis n. sp.

(Fig. 5; Pl. 6, figs. 3, 4)

HOLOTYPE: No. 27233, a left ramus with parts or all of I–M_3, collected by J. A. Dorr, July, 1950.

HORIZON AND TYPE LOCALITY: Upper Paleocene, Dell Creek local fauna, Hoback formation. UM-Sub-Wy Locality No. 1, Dell Creek Quarry, Dell Creek road near U. S. Highway No. 187, Sublette County, Wyoming.

DIAGNOSIS: A carpolestid near *Carpodaptes hazelae* Simpson in size. P_4 lower in relation to its length and with a more gently rounded lateral profile than in *C. hazelae* or *C. aulacodon*. Crest of P_4 is S-shaped in occlusal view due to deep posterointernal excavation of tooth and bears 5 distinct cuspules on main blade in addition to a sixth distinctly separate, talonid cusp. There may or may not be an incipient lower anterior seventh cuspule. These cuspules are not in a straight anteroposterior line, but conform to the S-shape of crest of tooth. An internal cingulum present on P_4 and extending forward somewhat beyond midpoint of tooth. M_{1-2} longer than wide with a weak internal cingulum at base of trigonid of M_1.

DESCRIPTION OF HOLOTYPE: (Table 6) The most nearly complete specimen of a carpolestid lower jaw thus far described in the literature. Enlarged, semiprocumbent incisor is complete except for tip which

was lost in preparation, but has been restored in outline on the figure as it was on the specimen. The tooth is laterally compressed throughout its length. It is long anteroposteriorly at base, but narrows rapidly upward to a rather acute point. Anterior edge is rounded. A broad, sharp-edged groove begins at posterointernal base of incisor and extends upward along crown, rotating slightly in a clockwise direction toward posterior edge of tooth before reaching its tip. A broad low ridge occupies center of this groove throughout its length. Posterior lip of the groove forms a sharp carina along posterior edge of tooth. The incisor root is long and extends backward and inward to a point below and slightly internal to the root of P_3. Correspondingly, the tip of the incisor is directed anteroexternally.

TABLE 6.—DIMENSIONS (IN MM.) OF LOWER DENTITION OF THE HOLOTYPE OF *Carpodaptes hobackensis* N. SP.

Holotype U.M.M.P. No. 27233	I	C	P_3	P_2	P_4	M_1	M_2	P_4 to M_2
Greatest length	1.5	.7†	.6†	.4	2.3	1.4	1.4	5.2
Greatest width	.8			.5	1.2	1.3	1.3	
Internal height of crown including cingulum	2.9*	.4†	Slightly lower than canine	Probably about as in P_2	1.8	1.0	.9	

* Including tip now missing.

† Measurement approximate from matrix impression.

The canine (following the terminology of Simpson) is broken, but a portion of the vertical, cylindrical root remains and an impression of the remainder of the tooth is preserved in the matrix. Canine root terminates ventrally against incisor at a point about midway on root of latter. Crown of canine is low and button-like with an overhanging anterior projection. Its crown barely reaches as high as posterior crown base of incisor.

All that remains of P_2 is likewise a portion of its root, which was long and may also have terminated against that of the incisor, and a matrix impression of the crown. Tooth was apparently similar to canine, but somewhat lower and perhaps with a less pronounced anterior projection of the crown. Crown of P_3 is broken, but this tooth must have closely resembled P_2. Incisor, canine, and premolars are all closely appressed with no diastema.

The typically carpolestid P_4 is lower crowned in relation to its length and its apex is less acute and more gently rounded than in *C. hazelae* or *C. aulacodon*. The posterior slope of its crest is not as steep as in those species. There are 5 well differentiated and widely separated apical cuspules and a distinctly separate talonid cusp on P_4. A slight swelling on crest below and anterior to apical cusps suggests that an incipient seventh cuspule is present. Apical cuspules are more widely separated from one another than on previously described species of *Carpodaptes*, the most posterior located well down on posterior slope of crest near talonid cusp. The talonid cusp is low as in other species of *Carpodaptes* and well below level of the trigonid of M_1. Ridges from the serrations of P_4 extend part way down internal face of the tooth and for a shorter distance down external face. Crest of the tooth in the occlusal view is not a straight anteroposterior line, but is S-shaped. The very anterior part of the crest is deflected labially and posterior half curves around a deep posterointernal depression in face of tooth so that the talonid cusp is offset lingually from midline of tooth. A sharp posterointernal crest descends from talonid cusp for a short distance toward base of tooth then swings anteriorly to become a well-developed internal cingulum. This extends forward as far as middle of tooth and is especially strong beneath the posterointernal depression.

M_1 has an elongate shearing trigonid blade which does not exactly continue the shearing blade of P_4. Meta- and paraconid are equidistant anteriorly and posteriorly from the protoconid. A weak internal cingulum is present below para- and protoconid.

M_2 is as in the other species of *Carpodaptes*. M_3 is present, but too incomplete for description. M_{1-2} are longer than wide.

Lower jaw, as preserved in the holotype, is extremely deep at symphysis thus accommodating the large and long rooted incisor. Posteroventral part of the jaw, however, may have been abraded thus decreasing its depth posteriorly and somewhat exaggerating the relative robustness of the symphysis.

REFERRED SPECIMENS: No. 27253, a jaw fragment with part of a left P_4, and No. 27234, a badly

waterworn jaw fragment with P_4 (broken)–M_1, are without doubt conspecific with the holotype. The teeth are too badly broken and abraded to show significant structural details not already better seen on the holotype. No. 27234 does, however, suffice to show the presence of a large mental foramen below P_{3-4}.

DISCUSSION: Matthew and Granger (1921, p. 6) briefly diagnosed the genotype, *Carpodaptes aulacodon*, from a single fragmentary specimen. The remaining two carpolestid genera, *Carpolestes* Simpson and *Elphidotarsius* Gidley, were then unknown. Simpson (1935, p. 10–12) figured and fully described the type of *Carpodaptes aulacodon* and at the same time redefined the genus so as to exclude the other two genera. Subsequent description of *C. hazelae* by Simpson and the present consideration of *C. hobackensis* necessitated further modification of the generic concept to the form given above.

The principal distinctions between *Carpodaptes* and *Carpolestes* are here considered to be: the relative size of P_4, the height of its crown, and the height and distinctness of the talonid cusp on that tooth; the length of the shearing blade and the relative positions of cusps on the trigonid of M_1; and character of the third lobe of M_3. The number and distinctness of the cusps on P_4 in *C. aulacodon* is apparently not characteristic of the genus, since *C. hazelae* has 5 distinct apical cusps, and the new species here described has possibly 6 cusps on the main blade.

Although the crest of P_4 and cusps thereon formed a straight line in both earlier described species of *Carpodaptes*, this is not considered a generic character of *Carpodaptes* since the same condition is present in two of the three species of *Carpolestes*. The fact that the third species of that genus, *C. dubius* Jepsen, has an S-shaped crest is not considered as indicating a close relationship between it and *Carpodaptes hobackensis*, but is taken as an indication that an S-shaped crest was developed in both *Carpolestes* and *Carpodaptes*. Hence it is considered permissible to assign *C. hobackensis* to the genus *Carpodaptes* in spite of the S-shape of its crest.

The possibility that the relative distances of the para- and metaconids anteriorly or posteriorly from the protoconids on the trigonid of M_1 might serve as a further distinction between the two genera was considered. Examination of a number of specimens proved this to be correct. In *Carpodaptes*, the two cusps are equidistant from the protoconid; in *Carpolestes*, the metaconid is closer to the protoconid than is the paraconid.

Carpodaptes hobackensis is placed in that genus on the basis of the characters discussed. It is distinguished from *C. aulacodon* and *C. hazelae* principally on the basis of its relatively lower, more rounded P_4 with an S-shaped crest and with an additional apical cusp, and by the fact that M_{1-2} are longer than wide in *C. hobackensis*.

Certain structural resemblances between *C. hobackensis* and different species of *Carpolestes* suggest that the former may lie in a more directly intermediate position between *Elphidotarsius* and *Carpolestes* than do either of the other two species of *Carpodaptes*. An internal cingulum on P_4 is as in *Carpolestes nigridens* Simpson; the S-shaped crest on P_4 as in *C. dubius*. The relatively high posterior portion of the crest and the increased number of cuspules on P_4 more closely approaches the general condition in *Carpolestes*. Also M_{1-2} are longer than wide in *Carpodaptes hobackensis* as is the condition in *Carpolestes*. Whether this more intermediate structural position has any stratigraphic significance is debatable. The possibility exists that *Carpodaptes hobackensis* is slightly older than the other known species of the two later Paleocene genera. Certainly it is no younger.

Primates, family indet.
(Pl. 5, figs. 10, 11)

A right $M_{1?}$ of a primate, No. 27521, cannot be identified with certainty, but compares closely with that of *Phenacolemur* Matthew in size and structure. The 3-cusped trigonid is small and high in relation to the low, broad, bicuspid talonid.

Another right $M_{1?}$, No. 27522, is much smaller than the specimen described above, but agrees with it in structure. It is smaller than M_1 in *Phenacolemur*, but larger than in a closely related form, from the Rock Bench of Wyoming, in the Princeton collection.

Order CARNIVORA
Suborder CREODONTA
Family ARCTOCYONIDAE
Thryptacodon australis Simpson, 1935
(Pl. 7, figs. 1, 2)

Thryptacodon australis SIMPSON, 1935. Am. Mus. Novitates, no. 817, p. 20–22, figs. 8–9.

Prior to 1947, no identifiable fossil vertebrates were reported from the early Cenozoic deposits of the Hoback Basin. In the summer of that year, the writer found a part of a creodont jaw at the horizon and locality from which the Dell Creek local fauna described here was subsequently quarried. The specimen, No. 24315, consists of a well-pre-

served fragment of a left ramus including the alveoli of P_4-M_1, a relatively unworn M_2, the alveolus of M_3, and the lower part of the coronoid process back to but not including the condyle. Dr. G. G. Simpson of the American Museum of Natural History identified the specimen as pertaining to the above genus and species. The writer reported its occurrence, the second specimen thus far known that unquestionably belongs to this species, before the Michigan Academy of Science, Arts, and Letters in the spring of 1948. It was suggested at that time that the Hoback formation was at least in part late Paleocene in age, but the report was unpublished. Horberg *et al.* (1949, p. 190) mentioned this "early carnivore jaw" in support of their suggestion that "at least the lower half of the Hoback formation is Paleocene."

The specimen was again compared with the holotype, A.M.N.H. No. 17384, in the course of this study. The 2 specimens are identical in the proportions and structural details of those portions of each which are preserved in common. They differ insignificantly in size, the Hoback specimen being very slightly larger.

On the lingual side of the jaw of the Hoback specimen is a broad and relatively deep groove. This groove extends from P_4 backward along the lower half of the jaw onto the ascending ramus. It is widest posteriorly, occupies half the depth of the jaw below M_2, then narrows rapidly anteriorly. A similar groove is present on the holotype and evidently did not result from crushing. Dimensions of M_2 are: length, 7.1 mm.; width, 5.6 mm.

Suborder FISSIPEDIA

Family MIACIDAE

Genus *Didymictis* Cope, 1875

Didymictis COPE, 1875. U. S. Geog. Survey, W. 100th Mer. (Wheeler), p. 5, 11.
Viverravus WORTMAN (and MATTHEW) 1899. Am. Mus. Nat. Hist., Bull., vol. 12, p. 136. [*non Viverravus* MARSH, 1872]
Didymictis MATTHEW, 1915. Am. Mus. Nat. Hist., Bull., vol. 34, p. 17–19.

Didymictis dellensis n. sp.
(Fig. 6; Pl. 7, figs. 3, 4)

HOLOTYPE: No 27232, University of Michigan Museum of Paleontology. A nearly complete right ramus with C–P_1 and P_3–M_2. Collected by Thomas Oelrich of the University of Michigan field party, summer of 1950.

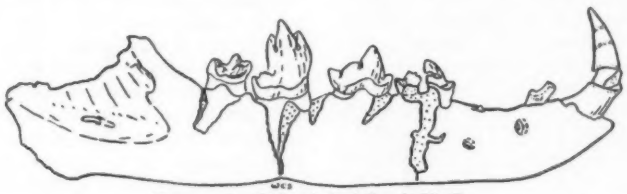


FIGURE 6.—*Didymictis dellensis* N. SP.
Holotype, U.M.M.P. No. 27232. Labial view, right ramus with C–P_1 and P_3–M_2, ×1. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming.

HORIZON AND TYPE LOCALITY: Upper Paleocene, Dell Creek local fauna, Hoback formation. UM-Sub-Wy Locality No. 1, Dell Creek Quarry, Dell Creek road near U. S. Highway No. 187, Sublette County, Wyoming.

DIAGNOSIS: Near *Didymictis protenus* (Cope) in size, but distinguished from the various subspecies of that species by the small size of P_1; dimensions, relative shortness of talonid, and elevated trigonid of M_2; and larger talonid cusps on P_{3-4}.

DESCRIPTION OF HOLOTYPE: No lower incisors preserved on specimen. Canine semi-procumbent at base, but, in spite of some exaggeration due to crushing, strongly recurved at tip. It is slender, somewhat laterally compressed with an internal groove and posterior carina. P_1 is single-rooted, reduced and trenchant, with a very small talonid cuspule. P_2 is missing, but was double rooted and apparently intermediate between P_1 and P_3 in size. Although broken, P_3 is trenchant, double-rooted, with a small anterobasal cuspule, a distinct posterior accessory cusp, and a broad well-developed crested talonid cusp. P_4 is trenchant, double-rooted, with a small but distinct anterobasal cusp, a large posterior accessory cusp, and a large distinct shearing talonid. There is a complete external cingulum.

M_1 has a sharp cusped trigonid. Protoconid is highest and inclines slightly posteriad. Paraconid slightly exceeds the metaconid in height. There is a low, short, basined talonid with hypoconid and hypoconulid separate and both higher than entoconid.

M_2 is reduced. Trigonid is equilateral, slightly elevated, and bears three distinct cusps. Talonid

is basined with a separate hypoconid, and confluent entoconid and hypoconulid. Talonid is short, being only very slightly longer than the trigonid.

The mandible is closely comparable to that of the Red Fox (*Vulpes*) in size and proportions. The ventral edge of the jaw, however, is nearly straight anteroposteriorly. The masseteric fossa does not extend to the ventral edge of the jaw, failing this by approximately 6 mm., and its lower border is nearly horizontal. The anterodorsal edge of the fossa projects slightly backward to form an overhanging lip. There are three mental foramina, one beneath posterior edge of P_1, another beneath the posterior root of P_2, and the third beneath P_4. The symphysial scar is an anteroposteriorly elongate oval extending backward to below midpoint of P_3. Dimensions of the holotype are given in Table 7.

TABLE 7.—DIMENSIONS (IN MM.) OF U.M.M.P. No. 27232, HOLOTYPE OF *Didymictis dellensis* N. SP.

	P_1	P_2	P_3	P_4	M_1	M_2
Greatest length	2.9	alveolar 6.2		9.0	10.2	6.6
Greatest width	1.6		Ca. 3.0	3.8	6.9	4.0

Alveolar length—C–M_2 61.2 mm.
Depth of jaw below P_4 15.0 mm.

REFERRED SPECIMENS: Four carnivore teeth, No. 27526, found within 3 inches of one another in the quarry, probably pertain to this species. These include left P^{3-4} which were in contact in the matrix. P^3 is trenchant and has a well-developed talonid cusp and a posterior accessory cuspule. The anterobasal cusp, if originally present, has been broken off. Metacone and metastyle are broken from P^4. The two remaining teeth are too badly shattered for description.

DISCUSSION: *Didymictis dellensis* is a larger form than *D. haydenianus* Cope and further differs from that species by possessing a posterior accessory cuspule on P_3 and a single such cuspule on P_4. Trigonid on M_2 is somewhat lower, but talonid is short as in the Torrejonian species.

Size alone suffices to distinguish *D. dellensis* from the much smaller Middle Paleocene *D. microlestes* Simpson and *D. tenuis* Simpson; likewise from the distinctly larger subspecies of the Eocene *D. altidens* (Cope).

The principal distinction between *D. dellensis* and the various subspecies of *D. protenus* (Cope) lies in the fact that the talonid is shorter in relation to the trigonid on M_2 of the former. (The same is true of *D. dellensis* as compared with *D. haydenianus*, but to a lesser degree.) An attempt was made to subject this character to measurement and numerical expression. Talonid length was measured from the most posterior extension of the crown of M_2 anteriorly to the point where the crest running anterointernally from the hypoconid intersects the posterior face of the trigonid. Thus measured, the talonid-trigonid length ratio of M_2 in *D. dellensis* is 1.06; in three American Museum specimens of *D. haydenianus*, 1.13; in five American Museum specimens, from the Clark Fork of *D. protenus proteus* Simpson, 1.51 (observed range, 1.31–1.70); and in one specimen of *D. p. protenus* (Cope), 1.48. The trigonid of M_2 in *D. dellensis* is slightly more elevated than in *D. protenus*.

Utilizing Simpson's (1937a, p. 16) analysis of *D. p. proteus*, the length of M_2 in *D. dellensis* is seen to be significantly shorter (13.3 times the standard deviation of the Clark Fork *Didymictis* from the mean of that group). The differences in width of M_2 (2.85 times the standard deviation) and of length-width ratio (1.80 times the standard deviation) are not conclusively significant.

Compared with *D. p. curtidens* (Cope) from the Sand Coulee and Gray Bull Eocene deposits, the talonid of M_2 is shorter, P_1 relatively smaller, and the talonid cusp on P_{3-4} stronger and more trenchant in *D. dellensis*. The overall dimensions of M_{1-2} in *D. dellensis* do not deviate significantly from the means of those teeth in *D. p. curtidens* analyzed in the paper cited above. Statistical data for the length of P_4 in *D. p. curtidens* (not given by Simpson) have been calculated for a series of 16 specimens in the American Museum collection as follows: observed range, 10.0 mm.–12.4 mm.; mean (length), 10.94 ± 0.17; standard deviation, 0.68 ± 0.12; coefficient of variation, 6.21 ± 1.1. The length of P_4 in *D. dellensis* (9.0 mm.) lies 2.86 times the standard deviation below the mean of that tooth in *D. p. curtidens;* possibly but not certainly a significant deviation.

D. dellensis appears to be an intermediate form between *D. haydenianus* from the Torrejon and *D. protenus proteus* from the Clark Fork.

Order CONDYLARTHRA

Family HYOPSODONTIDAE

Genus ***Haplaletes*** Simpson, 1935

Haplaletes SIMPSON, 1935. U. S. Nat. Mus., Pr. vol. 83, no. 2981, p. 243–44

Haplaletes diminutivus n. sp.

(Fig. 7; Pl. 5, figs. 12, 13)

HOLOTYPE: No. 27231, University of Michigan Museum of Paleontology, a left P^4–M^3, collected

TABLE 8.—DIMENSIONS (IN MM.) OF *Haplaletes diminutivus* N. SP. HOLOTYPE, U.M.M.P. No. 27231

	Greatest length	Greatest width	Ratio—length divided by width	Greatest external crown height including cingulum
P^4	1.8	2.4	0.75	1.3
M^1	1.9	2.6	0.74	1.2
M^2	1.9	3.0	0.64	1.1
P^4–M^2	5.6			

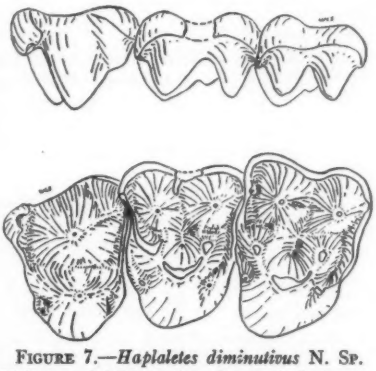


FIGURE 7.—*Haplaletes diminutivus* N. SP.
Holotype, U.M.M.P. No. 27231. Labial and occlusal views, left P^4–M^2, ×10. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming.

DIAGNOSIS: Smaller than *H. disceptatrix* Simpson, and lacking metacone on P^4. Protocone-hypocone separation on M^{1-2} not as distinct as in that species, hence internal bases are not so markedly bilobate. Named in reference to the small size of this species.

DESCRIPTION OF HOLOTYPE: P^4 is transverse. Paracone is high and conical with its rounded anterior edge distinctly separate from the strong and distinct parastyle, but with its sharper posterior edge confluent with metastyle. There is no indication of a metacone or of a conule. Protocone is slightly larger but lower than paracone and has an anterior and posterior carina. Anterior and posterior cingula extend slightly internal to protocone.

M^{1-2} are low crowned and bunodont. Both are wider than long although M^1 is less transverse than M^2. They have small proto- and metaconules united to anterior and posterior wings of protocone. There is a distinct although smaller hypocone posterior to protocone. Anterior, posterior, and external cingula are complete on molars. Anterior cingulum on M^{1-2} terminates abruptly internally in a slight cusplike enlargement on a line with protocone, but lower than and not joined to tip of that cusp. There are no external styles on these molars.

DISCUSSION: This is one of the smallest condylarths known. It approximates *Litolestes* Simpson in size, and is markedly smaller than *Haplaletes disceptatrix*, but is structurally much closer to the latter. Aside from the discrepancy in size, *H. diminutivus* differs from *H. disceptatrix* only in its lack of a rudimentary metacone on P^4 and in its slightly less bilobate internal molar bases. M^{1-2} are less transverse than those of *Litolestes*, but *H. diminutivus* has a general resemblance to *L. notissimus*. The two genera may be more closely related than Simpson (1937b, p. 228) suggests. Hitherto *Hapaletes* has been reported with certainty only from the Lebo beds of Montana although material

TABLE 9.—GEOLOGIC RANGES OF THE GENERA FROM UM-SUB-WY LOCALITY NO. 1

	Paleocene					Early Eocene	Later
	Puercan	Dragonian	Torrejonian	Tiffanian	Clarkforkian		
Allognathosuchus	×	×	×	×	×	×	Olig.
Haplodontosaurus			×	×			
Ptilodus		×	×	×			
Ectypodus			×	×	×	×	
Leptacodon				×			
Plesiadapis				×	×	×	
Carpodaptes				×			
Thryptacodon				×	×	×	
Didymictis			×	×	×	×	
Haplaletes			×	?			

by the University of Michigan field party, summer of 1948.

HORIZON AND TYPE LOCALITY: Upper Paleocene, Dell Creek local fauna, Hoback formation. UM-Sub-Wy Locality No. 1, Dell Creek Quarry, Dell Creek road near U. S. Highway No. 187, Sublette County, Wyoming.

from the 4350-foot level, high in the Melville beds of the Crazy Mountain field of Montana, has been questionably referred to this genus (Simpson, 1937b, Table 3, p. 48; p. 51).

Analysis of the faunule.—The invertebrate collection from this quarry has not been described, but Schultz (1914, p. 70) records a collection of invertebrates which includes similar forms from approximately the same horizon about 2½ miles southwest of the Dell Creek Quarry. He observes that, "These pureiy fresh-water fossils suggest the Evanston formation or so-called Upper Laramie." The Schultz collection was made from three localities, all in nearly the same horizon and very close together. All three together are equivalent to UM-Sub-Wy Locality No. 12 of this study.

The vertebrate faunule (Table 9) and the evidence for the age and correlation of the beds at the Dell Creek Quarry is as follows:

REPTILIA
- Crocodilia
  - Alligatoridae
    - *Allognathosuchus* sp. indet. A common genus ranging lower Paleocene to Oligocene.
- Squamata, family indet.
  - *Haplodontosaurus excedens?* Genus and species known from Torrejonian—upper Lebo beds of Montana and Tiffanian-Silver Coulee beds of Wyoming.

MAMMALIA
- Multituberculata
  - Ptilodontidae
    - *Ptilodus? fractus* n. sp. Genus hitherto reported from Dragonian (Dragon local fauna), Torrejonian (Torrejon, Rock Bench, upper Lebo local faunas), and Tiffanian (Melville fm.)
    - *Ectypodus powelli*. Genus known from Paleocene-Torrejonian to lowermost Eocene Gray Bull. Species hitherto known only from Tiffanian-Silver Coulee local fauna.
- Insectivora
  - Leptictidae
    - *Leptacodon packi*. Genus restricted to Tiffanian (Bear Creek, Melville, Silver Coulee local faunas). Species hitherto known only from Silver Coulee local fauna.
- Primates
  - Plesiadapidae
    - *Plesiadapis farisi* n. sp. Genus includes Tiffanian to Gray Bull representatives. Species shows closest relationship to Tiffanian species and in at least one respect, notably the strongly developed conules on P^{3-4}, may be less advanced than those forms.
    - *Plesiadapis* sp. A very distinct species from *P. farisi* in characters of P^4.
  - Plesiadapid, gen. and sp. indet. Much smaller than *Plesiadapis*. Near *Pronothodectes* in size and characters of upper incisors.
  - Carpolestidae
    - *Carpodaptes hobackensis* n. sp. Genus restricted to Tiffanian (Tiffany and Melville local faunas). This species may be slightly older than any previously known from the Tiffanian.
- Primates, family indet. Two isolated teeth comparing rather closely with *Phenacolemur*.
- Carnivora
  - Arctocyonidae
    - *Thryptacodon australis*. Genus ranges from Tiffanian to early Eocene. The species hitherto has been reported from the Tiffanian (Tiffany local fauna, and questionably from Melville fm.)
  - Miacidae
    - *Didymictis dellensis* n. sp. Genus ranges from Torrejonian to early Eocene. This species is intermediate between Torrejonian and Clarkforkian species.
- Condylarthra
  - Hyopsodontidae
    - *Haplaletes diminutivus* n. sp. Genus hitherto definitely known from the Torrejonian-upper Lebo beds of Montana and questionably from high in the Tiffanian-Melville beds of Montana. This species intermediate in some respects between *Haplaletes* and *Litolestes* (a small Tiffanian form).

Four previously described species are at least tentatively recognized at this locality. *Haplodontosaurus excedens* has been reported from both the Torrejonian and Tiffanian. The other three are known only from the Tiffanian.

Five new species are described. *Ptilodus? fractus* is of uncertain position. *Haplaletes diminutivus* may represent an intermediate stage between the Torrejonian species of that genus and the Tiffanian species of the genus *Litolestes*. *Didymictis dellensis* is apparently intermediate between Torrejonian and Clarkforkian species. *Carpodaptes hobackensis* and *Plesiadapis farisi* are certainly Tiffanian in their stage of development and may possibly be slightly older than the previously described Tiffanian species of those genera.

It appears evident that the beds at the Dell Creek Quarry are Tiffanian in age. The faunule is probably closest to the Silver Coulee local fauna from the Polecat Bench formation, but appears to be slightly older.

The unworn condition of the bone fragments

found in the quarry affords reason to suppose that the members of this faunule lived in an area close to the place where their remains are preserved (Case, 1936, p. 14). The occurrence of thin coal seams and carbonaceous shale above and below the limestone horizon from which the specimens were quarried suggests that swampy, heavily forested conditions constituted the environment in which the animals lived. The multituberculates, insectivores, primates, creodonts, and tiny condylarths were probably all forest dwelling and possibly in large part arboreal forms (Simpson, 1937b, p. 62–63).

UM-Sub-Wy Locality No. 7

Occurrence.—(Pl. 3; Pl. 4, fig. 3) The matrix here is a dark-brown, silty shale becoming gray upward. Calcareous, limonitic concretions are abundant. Two genera of mammals were recovered from this locality. The fossil remains are completely dissociated and mixed together.

Order PANTODONTA

Family CORYPHODONTIDAE

Coryphodon sp.

Coryphodon OWEN, 1840–1845. Odontography, p. 607, pl. 135, London.

Over 3 dozen dissociated teeth, No. 27255, from this locality indicate the presence of at least two individuals of this genus. The proximal half of a radius, No. 27256, also belongs to the genus and further work would undoubtedly result in the recovery of more nearly complete specimens. The material at hand is fragmentary and incomplete. Furthermore, the nature and extent of individual variation within the numerous species already described is poorly understood and the genus needs thorough restudy. In view of this, a more precise identification of this material seems neither justified nor worthwhile. The teeth compare closely in most respects with those of the holotype of *C. testis* (Cope), A.M.N.H. No. 4341. This species has been reported from the lower Eocene, Gray Bull of Wyoming and Almagre of New Mexico.

A fragment of an upper molar, No. 27531, of *Coryphodon* was found at Locality No. 16, at a stratigraphic horizon about midway between the Locality No. 7 and the Locality No. 2 horizons.

Order DINOCERATA

Family UINTATHERIIDAE

Probathyopsis successor Jepsen

Probathyopsis successor JEPSEN, 1930. Am. Philos. Soc., Pr., vol. 69, no. 4, p. 128–129, pl. 4, figs. 8–11.

Intermingled with the *Coryphodon* material were the dissociated remains of parts of two individual uintatheres, including two left lower jaws, one with a nearly complete series of cheek teeth. Upper and lower canines of two types; a number of isolated right and left upper molars and premolars; and several incisors were found which without much doubt belonged to one or the other of the individual uintatheres represented by the two jaws.

The Locality No. 7 specimens are identical in size and characters of the cheek teeth with the holotype, Princeton Museum No. 13234, of *P. successor* from the Gray Bull Eocene of Wyoming. Unfortunately, this species is only known from the holotype which does not include the lower jaw. The other two species of the genus, *P. praecursor* Simpson and *P. newbilli* Patterson, from the Clark Fork and Plateau Valley beds, respectively, are both based upon specimens which include the anterior portions of the lower jaws; neither holotype has a well-developed flange on the lower jaw. The larger of the two uintathere jaws, No. 27249 from Locality No. 7, is distinctly more advanced in flange structure than either of those Paleocene species. The smaller jaw, No. 27250, has the anterior end so broken that the nature of the flange is not determinable. There are two types of upper canines in the collection, one indistinguishable from that in *P. successor*, the other enormously enlarged, but both are definitely uintatherian in structure and in no way resemble the canines of *Coryphodon*, the only other form presently known to occur in this quarry. This evidence suggests two principal alternatives in the classification of this material: either (a) it affords evidence of extreme sexual dimorphism in *P. successor* (and possibly also in the other two species, the holotypes and only specimens of which could conceivably both be females), or (b) the material represents an entirely new genus (possibly including *P. successor*) which is conservative in the stage of its molar evolution, but has advanced appreciably in other characters. The first of these two alternatives has been selected as the more probable. It is quite possible that both *P. praecursor* and *P. newbilli* are based upon remains of females of those species. An extended description of this excellent material and further conclusions as to its systematic position are deferred with the expectation that further work proposed for the near future will add to the completeness of the collection. The stratigraphic significance of this find is already partially certain, however. The beds at Locality No. 7 are probably no older than those of the lowermost Eocene Gray Bull.

Analysis of faunule.—The genus *Coryphodon* occurs in deposits ranging from Clarkforkian to late Wasatchian in age (Van Houten, 1945, p. 436) and

is of little value in determining the age of the beds at this locality. The genus *Probathyopsis* is known from only a few specimens and those are all quite incomplete. The genus has been reported from the Tiffanian (Plateau Valley beds), the Clarkforkian, and lower part of the early Wasatchian Gray Bull beds. These specimens are referred to the earliest early Eocene species but, as discussed earlier, they might later prove to belong to a distinct and more advanced form. The age of the deposits at this locality and horizon is probably very earliest Eocene.

These two large archaic "subungulates" probably were members of a terrestrial or semi-aquatic flood plain facies rather than an arboreal faunal facies. Patterson (1939, p. 100-101) has suggested that the "feeding habits and diet of the hippopotamus are perhaps broadly comparable" to those of *Coryphodon*.

UM-Sub-Wy Locality No. 2

Occurrence.—The specimens at Locality No. 2 were all collected where they had weathered out upon the surface. The shale exposure from which they were collected occupies an area of very limited extent atop a ridge. As a result of solifluction, which ultimately brings the specimens to the surface, and trampling by deer and elk, who winter on this wind-swept ridge, the specimens are all extremely fragmentary.

Only one of the specimens collected was considered adequate for even tentative reference to a previously described species and none of the specimens adds anything to present biologic concepts of the group to which it belongs. The following systematic arrangement and description of this material is intended mostly as an aid to those who may later wish to re-examine the faunule for its bearing upon the stratigraphic position of the beds at this horizon, and the environment at the time of their deposition.

Order PRIMATES

Family ADAPIDAE

Pelycodus ralstoni? Matthew and Granger

(Pl. 7, figs. 5, 6)

Pelycodus ralstoni MATTHEW AND GRANGER, 1915. Am. Mus. Nat. Hist., Bull., vol. 34, p. 436.

A fragment of the posterior portion of a left lower jaw with an unworn left M_3, No. 27243, per-

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PLATE 4.—HOBACK FORMATION TYPE LOCALITY, PASS PEAK COBBLES, AND FOSSIL LOCALITIES

FIGURE 1. HOBACK FORMATION TYPE LOCALITY

Steeply eastward-dipping beds on Game Hill, along Hoback River and U.S. Highway No. 187, opposite mouth of Cliff Creek. Herein designated as type locality of Hoback formation.

FIGURE 2. TYPICAL "PRESSURE MARKED" QUARTZITE COBBLES COLLECTED FROM PASS PEAK CONGLOMERATE NEAR PASS PEAK

Note the healed fractures radiating from some of the pressure scars.

FIGURE 3. UM-SUB-WY LOCALITY NO. 1-DELL QUARRY HORIZON AND LOCATION

Indicated by white line and arrow. Black arrow indicates location of UM-Sub-Wy Locality No. 7—Uintathere Quarry on ridge in middle distance. Buildings are situated along U.S. Highway No. 187. Dell Creek Road (right center) joins U.S. 187 just left of highway bridge. Gros Ventre Range forms skyline.

PLATE 5—FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

(All figures ca. ×4)

All specimens on this plate are from the Dell Creek Quarry, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming.

	Pgs.
Haplodontosaurus excedens? (Gilmore)	74
Portion of a right dentary, U.M.M.P. No. 27217. (1) Labial view; (2) Lingual view	
Ptilodus? fractus Dorr n. sp.	74
Holotype, U.M.M.P. No. 27219. Left P_4–M_1. (3) Lingual view; (4) Anterior view of P_4.	
Ectypodus powelli Jepsen	77
(5) U.M.M.P. No. 27224. Lingual view. Left P_4.	
Leptacodon packi Jepsen	78
U.M.M.P. No. 27235, a broken right ramus in separate matrix halves, with portions of P_3–P_4. (6) Labial view of anterior half of jaw; (7) Lingual view of posterior half of jaw.	
Plesiadapis sp.	82
(8) U.M.M.P. No. 27519. Occlusal view of right P^4.	
Plesiadapid, gen. and sp. indet.	82
(9) Lingual view of upper incisor, U.M.M.P. No. 27520.	
Primates, family indet.	
(10) Occlusal view of a right M_1?, U.M.M.P. No. 27521	84
(11) Occlusal view of a right M_1?, U.M.M.P. No. 27522	84
Haplaletes diminutivus Dorr n. sp.	86
Holotype, U.M.M.P. No. 27231. Left P^4–M^2. (12) Labial view; (13) Occlusal view.	

FIGURE 1

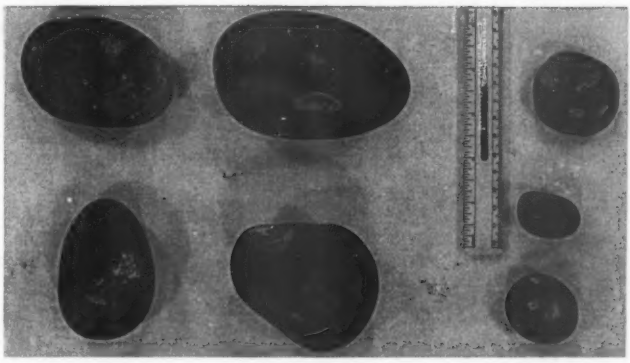
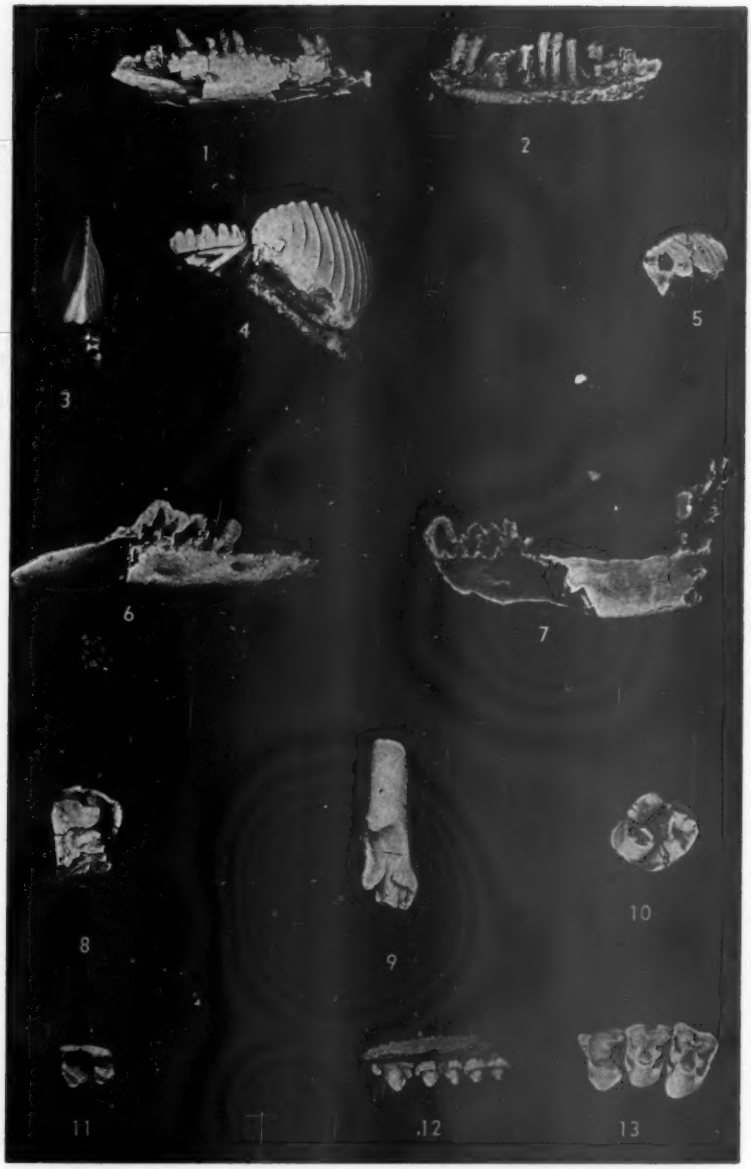


FIGURE 2

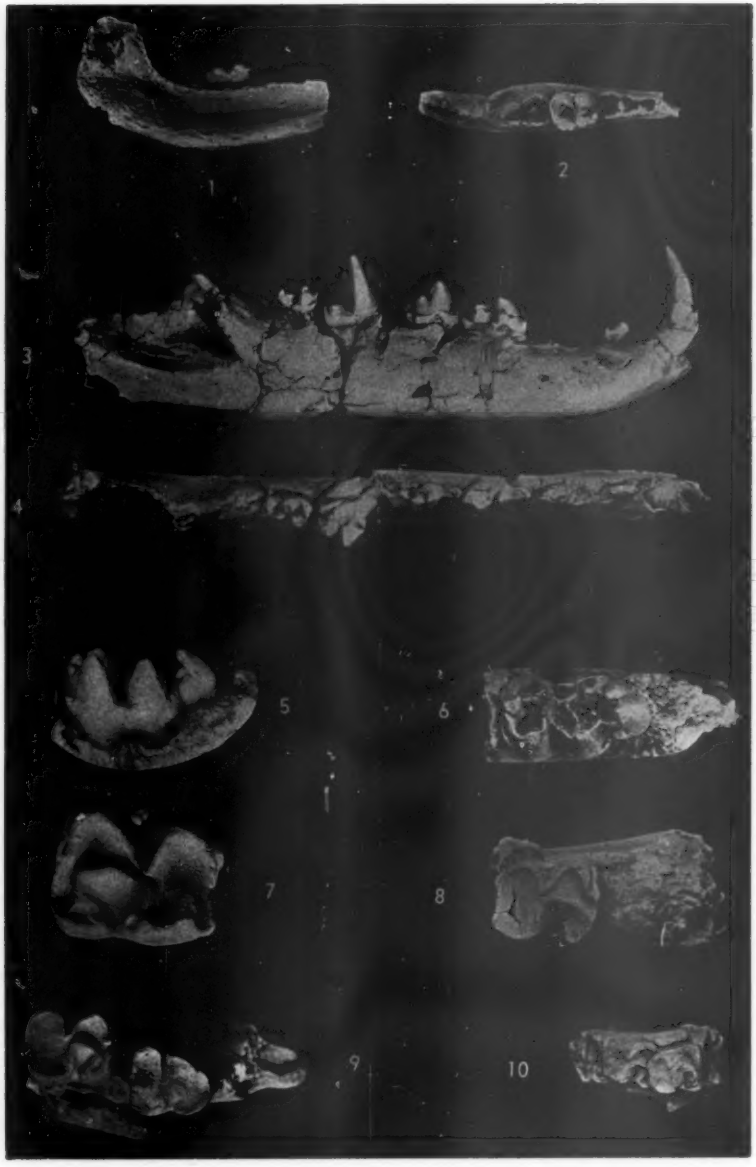
FIGURE 3

HOBACK FORMATION, TYPE LOCALITY, PASS PEAK COBBLES, AND FOSSIL LOCALITIES



FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

FOSSIL VERTEBRATES FROM THE HOBACK FORMATION



FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

tains to this genus and possibly to this species. There is a distinct paraconid on M_3, and the specimen agrees closely in other respects with the structure of that tooth in the type of *P. ralstoni*. M_3 is somewhat larger than in the holotype of the species, however, and the trigonid is slightly more elevated. Dimensions of M_3 are: greatest length, 5.5 mm.; greatest width, 3.4 mm.

Order TILLODONTIA

Family TILLOTHERIIDAE

Esthonyx sp. indet.

(Pl. 7, fig. 7)

Esthonyx COPE, 1874. Ann. Rept. Chief of Engineers, Wash. Append. FF, reprint, p. 1–18.

A left $M_{3?}$, No. 27244, is the sole specimen representing this genus in the collection. There is a reduced metastylid present. The specimen compares closely in size with a specimen labelled *E. burmeisterii* Cope, No. 16764, in the American Museum collection. Simpson (1937a, p. 5) has suggested that this species is probably a synonym of *E. bisulcatus* Cope. The specimen is significantly smaller, however, than specimens referred to *E. bisulcatus* in the American Museum collection. Dimensions of the specimen are: greatest length, 6.7 mm.; greatest width, 5.7 mm.

Order CONDYLARTHRA

Family HYOPSODONTIDAE

Hyopsodus sp. indet.

(Pl. 7, figs. 8, 9, 10)

Hyopsodus LEIDY, 1870. Acad. Nat. Sci. Philadelphia, Pr. vol. 22, p. 109.

Three specimens, No. 27240, a fragment of a right lower jaw with M_{2-3}; No. 27241, a right lower jaw fragment with $M_{1 \text{ or } 2}$; and No. 27245, a right lower jaw fragment with M_{1-2}, are referred to this genus, but are too poorly preserved and badly worn for specific identification. A fourth specimen, No. 27242, including a fragment of a lower jaw with worn right P_4–M_1, is similar in character but much smaller. It is also very slightly smaller than *Haplomylus speirianus* (Cope), but otherwise closely resembles the type of that species.

Order PERISSODACTYLA

Family EQUIDAE

Hyracotherium sp. indet.

Hyracotherium OWEN, 1840. Geol. Soc. London, Pr., vol. 3, p. 163.
Eohippus MARSH, 1876. Am. Jour. Sci., ser., 3, vol. 12, p. 401.

PLATE 6.—FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

Page

Plesiadapis farisi Dorr n. sp. 79
Holotype, U.M.M.P. No. 27230. A portion of a right maxillary with P^2–M^3. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming. (1) Labial view ca. ×4. Note infraorbital foramen above P^3; (2) Occlusal view. ca. ×4.

Carpodaptes hobackensis Dorr n. sp. 82
Holotype, U.M.M.P. No. 27233. A left ramus with parts or all of I–M_3. Hoback formation. UM-Sub-Wy Locality No. 1, Sublette County, Wyoming. (3) Lingual view. ca. ×10; (4) Occlusal view. ca. ×10.

PLATE 7.—FOSSIL VERTEBRATES FROM THE HOBACK FORMATION

Page

Thryptacodon australis Simpson 84
U.M.M.P. No. 24315. A left ramus with M_3. Hoback formation UM-Sub-Wy Locality No. 1, Sublette County, Wyoming. (1) Lingual view. ca. ×1; (2) Occlusal view. ca. ×1.

Didymictis dellensis Dorr n. sp. 85
Holotype U.M.M.P. No. 27232. A right ramus with Incisor and P_3–M_3. Hoback formation, UM-Sub-Wy Locality No. 1, Sublette County, Wyoming. (3) Labial view. ca. ×1; (4) Occlusal view. ca. ×1.

Pelycodus ralstoni? Matthew and Granger 90
U.M.M.P. No. 27243. A fragment of a left ramus with M_3. Hoback formation, UM-Sub-Wy Locality No. 2, Sublette County, Wyoming. (5) Labial view. ca. ×4; (6) Occlusal view. ca. ×4.

Esthonyx sp. indet. 91
(7) U.M.M.P. No. 27244. A left $M_{3?}$. Occlusal view. Hoback formation, UM-Sub-Wy Locality No. 2, Sublette County, Wyoming. ca. ×4.

Hyopsodus sp. indet. 91
(8) U.M.M.P. No. 27241. A right $M_{1?}$. Occlusal view. ca. ×4; (9) U.M.M.P. No. 27240. Right M_{2-3}. Occlusal view. ca. ×4.

Hyopsodontid, gen. and sp. indet. 91
(10) U.M.M.P. No. 27242. Right P_4–M_1. Occlusal view. ca. ×4.

Several isolated, deeply worn and broken teeth of this genus were found at Locality No. 2, and are included under Numbers 27237, 27238, and 27239. They are close to *H. borealis* Granger in size, but none of the specimens are sufficient for certain identification.

Analysis of faunule.—The following fossil mammalian genera and species have been identified from this locality:

Primates
  Adapidae
    Pelycodus ralstonii? Genus reported from early and middle Wasatchian. Species known from lower Gray Bull beds of Wyoming.
Tillodontia
  Tillotheriidae
    Esthonyx, sp. indet. (near *E. bisulcatus*). Genus known from Clarkforkian and Wasatchian.
Condylarthra
  Hyopsodontidae
    Hyopsodus sp. indet. An early and middle Eocene genus.
Perissodactyla
  Equidae
    Hyracotherium sp. indet. (near *H. borealis*) Genus restricted to the Wasatchian.

The age of the beds at Locality No. 2 thus appears to be early Eocene-Wasatchian and probably early Wasatchian.

The faunal facies is of the more normal open flood plain or savannah type (see Simpson, 1937b, p. 61–62). This is possibly, but, owing to the accidents of preservation and collection, not necessarily an indication that a change in over-all faunal facies occurred between the time of deposition of the beds at the Dell Creek Quarry and those at this locality. The abundance of hackberry seeds (*Celtis*) in this and higher horizons may indicate less humid conditions than those which prevailed at the time of deposition of the beds at the Dell Creek Quarry. Chaney (1925, p. 55) considered the occurrence of *Celtis* in Cenozoic deposits of western United States as indicative of a semi-arid or arid climate. However, he later modified this view (Chaney, 1936, p. 27) when he reported a floral assemblage which included *Celtis* from the late Cenozoic deposits of Kansas and interpreted that assemblage as indicating a mesophytic climate. This is certainly equivocal evidence in view of the fact that the same seeds are found at Locality No. 4 in association with the remains of aquatic vertebrates. The hackberry of today exhibits a wide range of environmental tolerance.

UM-Sub-Wy Locality No. 4

The beds at Locality No. 4 lie at the highest stratigraphic level in the Hoback formation from which even partially identifiable vertebrate remains were collected. The lithology and occurrence of fossils are similar to Locality No. 2.

A number of gar pike scales (*Lepidosteus*), No. 27532, several cranial fragments of an alligator, a few fragments of two turtles, and numerous seeds of the Hackberry, *Celtis*, have been collected at this locality (*Celtis* seeds are also common at Localities 2, 3, and 5). In addition, Mr. William T. Faris, during the summer of 1950, found a fragment of a left lower jaw with the posterior part of M_3, No. 27248. This specimen is questionably referred to the tillotherian genus *Esthonyx*. If this identification is correct, then the occurrence of this genus is an indication that the beds at this horizon are still Wasatchian in age.

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